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A CRITICAL REVIEW OF SUPERIMPOSED AND ANTECEDENT RIVERS IN SOUTHERN AFRICA

By

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ABSTRACT

In the past, various South African rivers have frequently but erroneously been described as "superimposed". By a strict interpretation of the definition, however, superimposed rivers are found limited to: portions of the primitive, interior drainage along which the rocks of the Karroo System have been removed and which now flow across underlying, older formations; and the extended-consequent, lower reaches of many rivers, which may display superimposition from the raised marine deposits of Cretaceous and Tertiary age fringing the continent. While the "superimposed poorts", by which the north-south flowing rivers break through the South Cape fold ranges, show little evidence of the origin advocated by such a description, they appear to owe their formation chiefly to the action of head-water erosion processes, subsequently controlled by master joint-systems.

Two types of antecedent rivers are recognised in South Africa. On the one hand, there are rivers which have not suffered dislocation in spite of the formation of disrupting structures across their courses. Most of this group are antecedent towards the late-Cretaceous and late-Tertiary crustal warpings, but a few show antecedence to Gondwanide diastrophism. On the other hand, some rivers are possibly antecedent towards late-tertiary desiccation and the resultant adverse conditions brought about by the spreading across their courses of a porous cover of Kalahari Sands.

Rivers showing combined superimposition and antecedence appear to be restricted to those which breach the highlands encircling the Bushveld Basin of the Central Transvaal.

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I INTRODUCTION

The descriptive adjectives, "superimposed" and "antecedent", have seldom been correctly applied by writers describing South African rivers. To the earlier recorders, they conveniently accounted for the abnormal river development often observed, and their over-frequent and inaccurate application is found in many papers. Later writers, with their better understanding of the past physiography of the country, observed the errors of their predecessors, and, realising the hazards accompanying the incautious interpretation of these terms, fought shy of them. They, therefore, often resorted to lengthy descriptions rather than risk the use of these simple, convenient terms. As a result, a literature has been built up in which either the terms, "superimposed" and "antecedent", recur frequently though in an incorrect sense, or these adjectives are rarely mentioned.

Being essentially descriptive of the development of the rivers, their correct use can only follow definite conceptions of the evolution of both the ideal river and the actual river described. It is on the basis of such a fundamental conception that this paper is written.

II THE PLACE OF SUPERIMPOSITION AND ANTECEDENCE IN THE IDEAL RIVER CYCLE

The position of every valley in a newly initiated drainage system is determined by irregularities in the surface slope or by inequalities in the resistance of the underlying rock. Valleys of the first group are usually termed "consequent", while those of the second fall under the broad heading of "subsequent". If, during their normal cycle of evolution from youth to old age, the rivers flowing in these valleys do not experience a disrupting influence from outside, the relation between the courses of these rivers and the factors controlling their original distribution and orientation remains clearly visible. During youth the impress will be strong and characteristic, but its influence will tend to diminish as maturity is approached. Finally, when the rivers have reached the ideal development of old age, this genetic relationship will have disappeared almost completely.

External factors which tend to disrupt this ideal evolution may result in the appearance, along a portion of a river's course, of structures quite foreign to those on which it originated. These structures could be formed after the establishment of the river, or they may be older structures, buried beneath the surface-cover on which the river took its rise and later exposed by the erosion processes of the river itself. Should their attempt to disrupt the established flow of the river succeed, then a new drainage will result which is in harmony with the new structure. If, on the other hand, the disturbing influence of the new structure is not sufficiently powerful, the river will be found to cut across this structure in such a manner as to suggest that its course cannot be genetically related to the foreign structure it now traverses. Rivers of this type may be divided into two main groups: *antecedent rivers*, when the foreign structures were formed after the establishment of the rivers; and *superimposed rivers*, when the foreign structures are older than the rivers.

A feature common to both antecedent and superimposed rivers is, therefore, that they both show departures from the normal, ideal cycle of river-evolution by virtue of their transgression of structures, foreign to their genesis. Differentiation of the two types is based on the comparison of the ages of the rivers and the foreign structures.

III ANTECEDENT RIVERS

The following definition of an antecedent river is based on that proposed by D. Johnson (1932, p. 493).

Any river, developed in response to any given geological structure, and across whose course there is developed some new and extraneous structure, is an antecedent river if it maintains an uninterrupted course across this structure. The new structure may be a gradually growing anticline, a slowly down-warped basin, or an uplifted fault-block. A river sufficiently vigorous to maintain its course across such an obstacle is then described as being "antecedent" to the formation of that obstacle.

The new structure should definitely constitute an obstacle to the continued flow of the river and should, ideally, have the effect of bringing the upstream portion of the river nearer its base level of erosion, while, downstream, the river may or may not be rejuvenated. Simple tilting of a land surface in the same direction as the flow of the river could then not be regarded as sufficient cause for labelling a river as "antecedent", as its only consequence will be an invigoration of the river's energies. Tilting in the opposite direction may, however, be the cause of an antecedent drainage. Although these latter streams are often classified separately as "anaclinal streams", (A. K. Lobeck, 1939, p. 173), they fulfill our conditions for antecedent streams and should not be regarded as a special, separate type.

IV SUPERIMPOSED RIVERS

It has already been mentioned that a "superimposed" river, or, more shortly, a "superposed" river, is one traversing an extraneous structure, which was in existence before the birth of the river but which has in no way guided the river's initial development. It is implied that the river acquired its course in harmony with the constructional surface or the internal structure of an overlying formation, under which the foreign structure was buried. By cutting down through, or stripping, these overlying deposits, the river has had its course "imposed from above" and is, therefore, called a "superimposed" river.

The genetic relationship between the rivers and the overlying beds may act as a basis for the distinction between different types of superimposed rivers. Thus we may describe the lower reaches of some of the South African rivers as being "superimposed extended-consequent" rivers, implying thereby that they originated as extended-consequent rivers on an overlying surface, and have since been superimposed on to an unconformable structure.

With the notable exception of D. Johnson, on whose writings the above definition has been based, most writers appear to have too narrow a concept of superimposition. Thus Grabau (1920, p. 744) and von Engeln (1942, p. 224) require that the covering formations should be "deposits of coastal plain strata", which would result in the formation of only superimposed consequent, and superimposed extended-consequent rivers. A. Holmes (1944, p. 182) on the other hand, seems to limit the

underlying foreign structure to "ancient folded rocks" though it is possible to picture a drainage system, in harmony with the internal structure of, say, a region of folded rock, being let down on to an underlying igneous body with the formation of superimposed subsequent, superimposed resequent, superimposed obsequent, etc., rivers.

It also happens that some writers have broadened the term "superimposed" to describe rivers that have been let down upon lower parts of the very structure that determined their genetic pattern. This conception is so broad that it will necessarily include most South African rivers, and becomes meaningless. Following Johnson's proposal, these rivers should rather be referred to as "inherited rivers", since their courses have been inherited from higher levels.

Two important features are, therefore, necessary for the recognition of the superimposed character of a river: firstly, imposition of the river course from above; and secondly, imposition on to a foreign structure. Since the first of these requirements alone satisfies the requirements for inheritance, superimposition must be regarded as a special case of inheritance.

The foregoing discussion of superimposed and antecedent rivers may create the impression that an entire river is simply described as either a normal, a superimposed, or an inherited river. It is, therefore, perhaps necessary to point out that, owing to their extraordinary length and age, many rivers cross a variety of structural regions and the nature of their labours may vary from place to place, and from time to time. Along one stretch a river may be actively eroding, along another it may be alluviating; here it may show structural adaptation, while there it may be structurally transgressive. Because of this diversity it is, therefore, not unusual to find that rivers are only superimposed or antecedent along portions of their courses, and often stretches showing superimposition may overlap with stretches showing antecedence. On the whole, with long rivers, complex drainage is more often the rule than the exception.

V THE KAFFERKUILS RIVER

Before attempting a general examination of South African rivers with the purpose of determining the existence of any antecedent or superimposed rivers, a detailed investigation of single examples of an antecedent river and a superimposed river will illustrate the characteristic evolution of such rivers. The Kafferkuils River, draining part of the southern coastal region in the neighbourhood of Riversdale, has been chosen as particularly suitable for this purpose as both these forms are developed along portions of its course. It has been thoroughly examined and fully described by M. S. Taljaard (1949, pp. 82-86), and is shown to be antecedent to a late-Tertiary warp axis which cuts its mid-course roughly perpendicularly, and to be superimposed upon closely folded, east-west dipping Bokkeveld strata along its lower reaches.

There is reasonably clear evidence that the Kafferkuils River was in existence in early-Tertiary times. During the Eocene epoch, its headwaters must have drained the southern fault-scarp of the Langeberg range, while, lower down, it had succeeded in base-levelling a plain across steeply folded Bokkeveld strata. The main stream must have reached at least a mature age and its course across the coastal plain was meandering and migratory. Its chief tributaries, too, notably the Brak and Soetmelks Rivers, possessed meandering courses as they approached the main stream, but many of the smaller tributaries were still strongly influenced by the structural build of the region and their straight courses followed the east-west strike of the underlying Bokkeveld beds.

During this period, a gradual subsidence of the land was accompanied by an incursion of the sea at the margin of the continent, a process which was finally halted when the coastline in this region lay nearly 15 miles inland from its present position. During the first stages of this slow submergence, wave erosion replaced river erosion for a time sufficient to produce limited plaination, but too brief to obliterate the courses of either the main streams or their tributaries. Continued subsidence lowered this surface below the base level of wave erosion and a blanket of marine deposits, at least 350 feet thick, was laid over this surface during the Oligocene and early-Miocene epochs. Inland, the river was simultaneously aggrading a flood-plain composed of a veneer of terrestrial gravels.

General, differential uplift of the whole of the South African sub-continent occurred in the late-Miocene epoch. Along this fringe of the continent the uneven nature of the upheaval was responsible for the formation of a warp-axis, running approximately east-west and cutting the Kafferkuils River just north of Riversdale. This warping was not the result of extra uplift along a line, but was caused by the lagging behind of the regions to the north and south of the axis. Although the uplift itself may have been rapid, the warping must have been a fairly slow process, and downward erosion by the river was sufficiently active to keep pace with the formation of this obstacle. The attempt to disrupt the river was consequently defeated and the river is antecedent to the warping. Where the river crosses this warp axis, its course has been deeply incised. On the down-stream side, both the increased gradient due to tilting and the general lowering of base-level have contributed towards the rejuvenation and deep incision of the river. The stretches upstream from the axis have, however, been tilted against the general flow and a certain amount of ponding was experienced by both the Kafferkuils River and its tributary, the Vet River. To overcome this gradual loss of gradient, the rivers have built up their courses by aggradation and they now possess low-sloping, swampy stream-beds, built on their own alluvium. The head-waters to the far north remain youthful and active, and are slowly lengthening their courses backwards into the Langeberg range. While the actual presence of the warp axis may be deduced from the flexed attitude of the formerly smooth line of the Oligo-Miocene river gravels, the antecedent relationship of the river to the warping is best shown by the rejuvenation below, incision across, and ageing above its intersection with the warp axis.

The mid- to late-Tertiary uplift was also responsible for lowering the relative position of sea-level, thereby exposing a broad stretch of the sea-floor to the influence of sub-aerial agencies. This stretch was composed of sandy limestones of Oligo-Miocene age, covering, unconformably, the bevelled, folded Bokkeveld strata, and its flat and featureless surface possessed a gentle seaward slope. The Kafferkuils River, which had emptied into the mid-Tertiary sea at a point just south of its junction with the Soetmelks River, had its course extended across this new, flat surface, in such a way that its relatively straight direction of flow was in harmony with the slope of this constructional surface. Therefore, like the straight lower reaches of so many other South African rivers, this portion of the Kafferkuils River may be genetically classified as "extended-consequent" across the uplifted sea-floor strata. The semi-consolidated nature of this sandy limestone offered little resistance to the attack of this rejuvenated river and its incision was consequently rapid. After cutting through 350 feet of the arenaceous limestone, the river encountered the underlying Bokkeveld strata, the internal structure of which was against the flow direction of the river. However, since the river was sufficiently active and firmly established, it was able to retain both the positional and the directional characteristics of its consequent course, in spite of

the deflecting impulsion exercised by the Bokkeveld strata. Below its junction with the Soetmelks River, therefore, the Kafferkuils River conforms in all respects to the definition of a superimposed river. Firstly, it was developed in harmony with the primary slope of a newly exposed constructional surface; and secondly, it now traverses and transects a re-exposed foreign structure in such a way that no genetic relationship is observed between the river's flow direction and the structure. This portion of the Kafferkuils River may, therefore, be termed a "superimposed extended-consequent" river. Incision into the Bokkeveld beds of 300 feet was measured by Taljaard, and the total incision of 650 feet is said to reflect the lowering of base level of erosion by an equal amount. This lowering was not only the result of mid- to late-Tertiary upheaval, but was also caused by a general eustatic lowering of sea-level in pre-Pleistocene times.

It is further shown by Taljaard that most of the tributaries along this superimposed portion of the Kafferkuils River have, by a process of head-water tapping, uniquely excavated their pre-Oligocene valleys, and their resurrected courses show a close alignment to the underlying Bokkeveld structure. The valley of the present Kafferkuils River, although it probably follows roughly the same general direction as that of the "old" Kafferkuils, cannot, however, be regarded as a resurrected river, as its straight course is mainly an expression of its extended-consequent channel, established for the first time on the covering surface of limestone, and is not apparently connected with the meandering channel of the "old" Kafferkuils River. The writer would like to suggest that portions of this buried, meandering course have been excavated by some of the tributaries, as this would seem to supply a satisfactory explanation for the frequent junction of two tributaries on opposite sides of the main river, in such a way that their courses are continuous and slightly transgressive towards the Bokkeveld structure.

The most recent alteration in the position of the coastline was caused by a eustatic rise of sea-level in the late-Tertiary epoch. The subsequent invasion of the ocean, which filled and drowned the lower reaches of the over-lengthened course of the Kafferkuils River, has in no way altered the superimposition or antecedence of portions of the river.

VI A BASIS FOR THE EXAMINATION OF SOUTH AFRICAN RIVERS

The examination of the Kafferkuils River system shows how, to a large extent, the interpretation of a river as either superimposed or antecedent is facilitated by a knowledge of the river's past history as regards both its origin and its relation to the relative movements of sea-level and to crustal deformation. An investigation of the rivers of Southern Africa, with the object of determining possible superimposition and antecedence, should, therefore, be preceeded by a study of the physiographic evolution of the sub-continent since the initiation of the current drainage. Such a study has been made and recorded by many writers, and a repetition and considered discussion of the theories advanced would here be unnecessary, unprofitable and out of place. Differing as these various opinions do, it is felt that one definite conception of the physiographic evolution of this country must be chosen on which to base the examination that is to follow. For this purpose, the views of A. L. du Toit and M. S. Taljaard, so concisely expressed in the final chapters of "Geology of South Africa" and "A Glimpse of South Africa" respectively, have been selected.

It is commonly agreed that, from the time of the eruption of the basaltic lavas of the Stormsberg Series, which marked the close of the Karroo cycle of deposition

in early-Jurassic times, to the present day, both Central and Southern Africa have continuously been land, with the exception of narrow marginal zones where encroachment of the sea occurred in Cretaceous and Tertiary times. The interior has, during this period, been subjected to many cycles of uplift, doming and warping, interspersed with times of crustal stability during which river systems were responsible for extensive erosion and even plaination. It seems likely, therefore, that much of our present-day drainage had its birth on this surface of basaltic lavas, uplifted in early-Jurassic times. In a search into the past of the Southern African continent to determine when and where conditions have been suitable for the superimposition and antecedence of rivers, it is, therefore, only necessary to examine conditions subsequent to the close of the Karroo cycle of deposition.

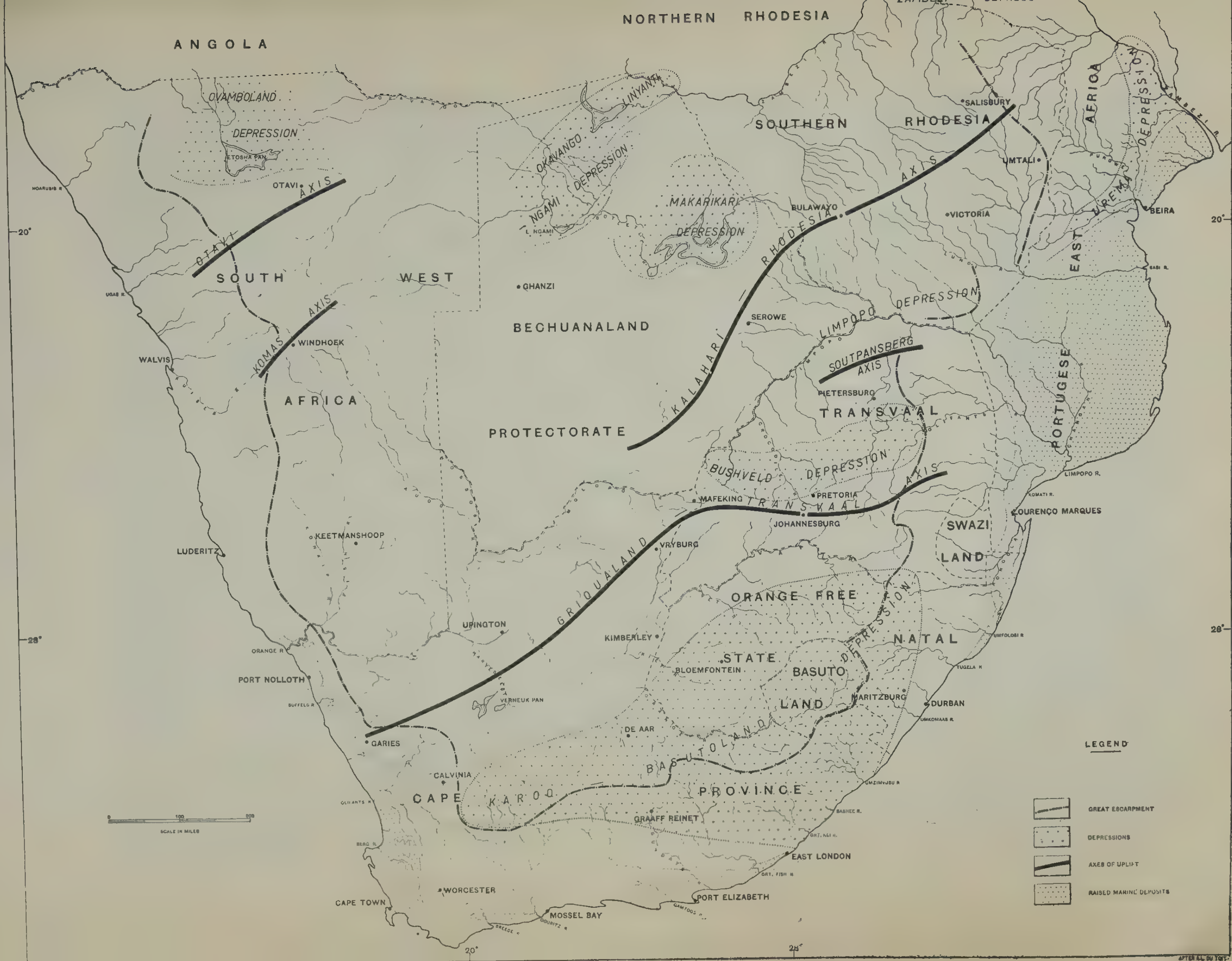
VII SUPERIMPOSITION IN SOUTHERN AFRICA

From the definition, it is clear that the first necessary condition for the superimposition of a river is that it should become established in harmony with the internal structure, or the surface slope, of one distinct stratigraphic unit, which bears an unconformable attitude to an older, underlying structure. In Southern Africa there exist three such unconformably-covering formations from which the present-day drainage lines could have been superimposed. They are the rocks of the Karroo System, the Kalahari Sands and the raised marine deposits of Cretaceous and Tertiary age

The second condition is that sufficient uplift to enable the rivers to lower their courses on to the underlying formations should have been experienced by the country, Uplift of varying magnitude has affected the crust on many occasions, and we will. therefore, examine the rivers draining the aforementioned formations and determine whether they have experienced sufficient uplift to lower their courses to a position of superimposition across the underlying structures.

(*a*) *Superimposition from the Karroo formations.*

The major rivers of the interior are all of considerable antiquity, their origin dating back to the close of the Karroo depositional cycle, when their courses were laid down on the uppermost surface of the Karroo rocks, in such a way that their patterns were in harmony with the consequental slope of this new constructional surface. At this time, Southern Africa formed the interior region of the super-continent, Gondwanaland, and the eastern drainage had been divided from that towards the west by the Lebombo monocline, which had been formed by crustal flexuring during the Liassic epoch. The primitive, interior drainage was also appreciably influenced by further, contemporary crustal deformation, notably the "Rift movements", described by du Toit (1939, p. 511), which imparted the north-eastern, or south-western, trend to the upper Orange River and some of its tributaries, the middle Limpopo River and the middle Zambezi River, etc., and the south-eastern trend to the lower Zambezi and the lower Limpopo Rivers. The form of these primitive rivers appears to have altered little and their present courses may be regarded as having been inherited from their Jurassic counterparts. In addition, they have, over long stretches, succeeded in uncovering pre-Karroo formations and their courses have been superimposed on such formations. Thus, the middle and lower Orange River, most of the Vaal River, the Limpopo, the Olifants, the Sabi and a small stretch of the Zambezi Rivers all possess courses that have been superimposed from the rocks of the Karroo System.



ANGOLA
NORTHERN RHODESIA
SOUTHERN RHODESIA
OVAMBOLAND DEPRESSION
ETOSHA PAN
OTAVI
OTAVI AXIS
SOUTH WEST AFRICA
KOMAS AXIS
WINDHOEK
WALVIS
LUDERITZ
KEETMANSHOOP
PORT NOLLOTH
ORANGE R.
GARIES
CALVINIA
CAPE PROVINCE
KAROO
CAPE TOWN
WORCESTER
MOSSEL BAY
PORT ELIZABETH
EAST LONDON
GRAAFF REINET
DE AAR
UPINGTON
VERNEUK PAN
GRIQUALAND
KIMBERLEY
BLOEMFONTEIN
ORANGE FREE STATE
BASUTO LAND
BASUTOLAND
NATAL
NATAL DEPRESSION
MARITZBURG
DURBAN
SWAZI LAND
LOURENÇO MARQUES
LIMPOPO R.
PORTUGESE EAST AFRICA
BEIRA
UMTALI
SALISBURY
VICTORIA
BULAWAYO
RHODESIA AXIS
KALAHARI
LIMPOPO DEPRESSION
SEROWE
SOUTPANSBERG AXIS
PIETERSBURG
TRANSVAAL
BUSHVELD DEPRESSION
PRETORIA
MAFEKING
TRANSVAAL AXIS
JOHANNESBURG
VRYBURG
BECHUANALAND PROTECTORATE
GHANZI
OKAVANGO DEPRESSION
NGAMI
L. NGAMI
LINYANTI
MAKARIKARI DEPRESSION
UREMA DEPRESSION
ZAMBEZI R.
20°
28°
SCALE IN MILES
0
100
200
LEGEND
GREAT ESCARPMENT
DEPRESSIONS
AXES OF UPLIFT
RAISED MARINE DEPOSITS
AFTER A.L. DU TOIT

Along these ancient rivers, the superimposed characteristics are not always equally apparent. The most obvious development is naturally to be expected where the uncovered foreign structures are such as exert a marked impression on the form of the land surface, and it is usually only across these structures that the rivers are described as being superimposed. Such structures have been exposed by the Olifants River in the Eastern Transvaal where it has been superimposed obliquely across the alternating hard and soft strata of the Transvaal System, cutting through hard and soft layers alike and showing little adaptation to the geologic structure of the rocks. Normally, the structural influence of this System on the drainage and the landscape is extreme, the former developing subsequent, strike streams and the latter a "banke-veld" topography. For the larger parts of their courses, however, these rivers have uncovered the relatively homogeneous granitic basement of the sub-continent and their superimposed attitude is not as clearly defined. The presence of the former covering of Karroo rocks is, however, generally revealed in the opposing valley flanks. These portions must also be regarded as superimposed, as their historic development has essentially paralleled that of the other portions.

Two further instances of the transection of pre-Karroo structures, for which the superimposition of these rivers from Karroo strata has been responsible, may be mentioned. Firstly, between Prieska and Buchuberg, the Orange River cuts across strata of the Transvaal and Matsap Systems, the superimposition probably having occurred from a surface lying at about 4,000 feet and continuous with the surface of the Kaap Plateau. This surface was apparently the "mid-Miocene Peneplane" of Dixey. A better display of superimposition is seen along the Vaal River in the vicinity of the Vredefort dome. Here the river follows a haphazard course, completely unrelated to the underlying structure, across the steeply upturned strata of the Transvaal, Ventersdorp and Witwatersrand Systems and Old Granite. The profile of this river shows that the lithological resistance of certain of these structures has had but little effect on the general gradient of the river (L. C. King, 1948, p. xxv). The existence of a formerly unbroken mantle of Karroo strata is disclosed by the many residuals of this cover which still persist more especially towards the south. An interesting feature of the Vaal River is that, except in its headwater region, its course coincides remarkably with the boundary between Karroo and pre-Karroo rocks and the river flows more or less on the re-exposed pre-Karroo surface. Cooke (1946, p. 245) accounts for this by the lateral migration of the river in a southerly direction, following the exposure of the pre-Karroo surface which had a definite southward slope. This migration is borne out by the fact that the Tertiary high-level gravel deposits are almost entirely confined to the right bank of the river. To what extent the river may still be regarded as superimposed after this migration, which is in fact an adaptation to the nature of the unconformity, will be considered later.

The middle courses of both the Limpopo and the Zambezi Rivers occupy valleys which coincide with gigantic troughs, the floors of which have sunk a couple of thousand feet beneath the plateau block on either side. Details of these lengthy depressions are scanty but they probably coincide with pre-Karroo troughs, in which vast accumulations of Karroo sediments collected and which have suffered subsequent post-Karroo sagging. The Limpopo River has succeeded in removing the Karroo strata along its valley and its course has been superimposed on to the pre-Karroo structures. The Zambezi River apparently occupies a deeper trough, from which it has not yet removed the Karroo rocks. Consequently, except for small stretches at the Kariba Gorge and between Chicoa and Tete, it flows almost entirely along strata of Karroo age, and only along these two small portions does it show superimposition.

Some of the tributaries entering the mid-Limpopo River from the south flow across the faulted structure of the Zoutpansberg highlands in a manner which suggests their superimposition across these structures from a former cover of Karroo rocks (L. C. King, 1942, p. 281). The existence of such a cover over this region is, however, unlikely as this region was topographically prominent in pre-Karroo and Karroo times. Thus, Taljaard (1949, p. 178) believes that the Zoutpansberg and the adjacent Pietersberg "became the stripping ground and source region for the sediments of Karroo age, and could hence not receive deposits of similar age". This naturally disposes of a superimposed origin for the poorts of the Magalakwin, Brak and Sand Rivers across the faulted structure of the Zoutpansberge. Rather were these rivers members of the northerly drainage of Karroo times and, by headwater erosion, succeeded in breaching the quartzite ridges at an early date. Extension into the regions behind the mountains proceeded apace with the stripping of this less resistant granite country, with the formation of the even surface of the Pietersburg Plateau.

So far, only the interior drainage of the continent, which shows inheritance from the primitive drainage of the Jurassic surface underlain by Karroo rocks, has been examined for superimposition from Karroo strata. Evidence, however, shows that these strata formerly extended over and beyond the Southern African coastal borders and that their counterparts are found on the adjacent fragments of Gondwanaland. Disregarding for a moment the drainage of the South Cape Fold region, there is a possibility that the present marginal drainage between the Great Escarpment and the sea originated on this "Karroo" surface and now shows superimposition from it. The present surface is not, however, an inheritance from the "Karroo" surface, but is a younger surface which has been cut back from the coast by the active headwater erosion of a new drainage, established in the Cretaceous Period and closely related to the disruption of Gondwanaland. The courses of these rivers were extended by the activity of their obsequent headwaters against a retreating scarp, and these headwaters are, at present, still situated along the scarp-front or just behind the scarp, which they cross by means of spectacular waterfalls. This is distinctly different from the method whereby the superimposed, primitive, interior drainage crosses the Great Escarpment.

A certain amount of post-Cretaceous uplift of this coastal margin has been responsible for lowering the rivers to some extent, and in places superimposition on a small scale is shown across exposed foreign structures, which may be in the form of buried fragments of the Primitive System, or may have resulted from post-Karroo faulting. An interesting example of the superimposition of a river across the latter type of structure is found at Port St. John's, and is described by du Toit (1920, pp. 36-41). Here, the extended-consequent course of the Umzimvubu River has been superimposed in a remarkable cleft passing undeflected through the "Port St. John's horst", a faulted block capped with resistant Table Mountain Sandstone and standing above the level of the surrounding country.

On the whole then, the larger rivers of the Southern African interior, by virtue of their inheritance from the primitive Jurassic drainage, are in a position to show superimposition from the Karroo cover, and this has occurred extensively wherever underlying, pre-Karroo structures have been exposed. A similar development is not, however, indicated for the more recent marginal drainage.

(*b*) *Superimposition from the Kalahari Sands.*

Towards the end of the Tertiary period, climatic conditions over the interior were semi-arid and a thick mantle of Kalahari Sand was spread by wind over large tracts of the continent. Especially in Southern Rhodesia and the Bechuanaland

Protectorate, much of the existing drainage was choked by this porous cover and only those larger rivers experiencing periodical floods were able to keep their channels clear. It is possible that, with a return to more humid conditions, a new consequential drainage system was established on this surface and that these rivers, by virtue of their establishment on the surface of a formation blanketing older structures, would then be potential superimposed rivers. Those rivers then, which had their origin on this cover and have since succeeded in removing these sands from their courses could then be regarded as superimposed rivers. If, however, the rivers follow drainage lines which were in existence before the spreading of the sands, they would be antecedent to the foreign structure constituted by the sands. The distinguishing criterion between the two types would be the age relation between the river and the deposition of the sands. Whereas in antecedence the sand cover would be younger than the river and would be regarded as the foreign structure, in superimposition the river would be younger and the underlying, buried structure would be foreign to its development.

Evidence on which to base this dating is scanty, and conflicting reports on the same feature are often noted in the writings of a single investigator. Thus, in one report, du Toit (1933, p. 14) regards the deposition of these Kalahari Sands as preceeding the late-Tertiary crustal warping, and the resultant consequent streams, thrown off at right angles to the axes of uplift, would represent a new drainage, established in harmony with the slope of the surface of the Kalahari Sands. On this basis, the tributaries of the Zambezi, the Limpopo, and the Sabi Rivers which take their rise on the flanks of the Kalahari-Rhodesia axis of uplift would have superimposed consequent courses from the surface of the Kalahari Sands along those stretches where they have succeeded in uncovering the basement rocks.

Elsewhere, du Toit (1939, p. 511) writes that: "the conspicuous sub-parallel rivers flowing south-east and north-west from the Rhodesia-Kalahari axis are manifestly 'consequent streams' resulting from upheaval along a line and incidentally *antedate* the Kalahari Sands." This latter report is probably more accurate, in which case a superimposed development could not be entertained for these rivers.

Although possible, it is unlikely then that any important superimposition of these rivers has occurred from the Kalahari Sands.

(*c*) *Superimposition from raised marine deposits of post-Karroo age.*

Uplift of the sub-continent and the exposure of the continental shelf deposits resulted from the adjustment of isostatic equilibrium during the late-Cretaceous and late-Tertiary periods. The distribution of such marine deposits about the margin of the continent is shown on the accompanying diagram.

The existing rivers indicate, by their meandering character, that, prior to the uplift, they developed on a plain having a very low gradient. During uplift, these serpentine courses were incised and their actual form has been well preserved. Nevertheless, in most cases their channels straighten out again before reaching the sea. This phenomenon is due to the extension of these rivers across the newly exposed sea-floor, in a direction perpendicular to the coast-line, as the latter migrated outwards. Where these rivers, therefore, cross the raised marine deposits, their courses may be described as extended-consequent, governed by the gentle seaward slope of the newly exposed deposits. Where uplift has been sufficient to lower the base level of erosion to a position below the base of the marine deposits, superimposition of these extended-consequent courses, along lines similar to those already described for the Kafferkuils River, is possible.

Without a doubt, many South African rivers do show this superimposition of their lower courses, just as many have not experienced sufficient uplift to be able to do so. Unfortunately, insufficient detailed investigations have been recorded to make possible an enumeration of those rivers which are superimposed extended-consequent and those which have not yet reached this stage of development.

(*d*) *The "superimposed" poorts of the Cape Fold ranges.*

An interesting and complex stream pattern has been evolved by the rivers draining the fold-belts of the Southern Cape, stretching from East London in the east to Worcester in the west. The east-west striking folds into which the rocks of the Cape System and the lower strata of the Karroo System have been flexed are well known. Topographically, the most prominent feature of the region is the two sub-parallel chains of mountains which result from the exposure of the crests of resistant anticlinoria of Table Mountain Sandstone, while in the synclinal valley between the two ranges smaller, humped mountains are built by pitching anticlines of Table Mountain Sandstone. On the whole, the trellis drainage pattern of this fold region is strongly controlled by the structural build of the crust, east-west flowing streams following the softer strata exposed in the synclines while their resequent tributaries branch off at right angles to drain the mountain flanks. The major rivers, into which this east-west drainage flows, have their headwaters along the Great Escarpment to the north, and, after flowing across the Great Karroo, they cut across the structural trend of the fold belt. This they do by traversing the mountain ranges by means of narrow, steep-sided defiles or "poorts". Most writers have explained the origin of these poorts as due to the superimposition of the north-south flowing rivers, and a casual glance at a map of this region confirms the attractiveness of this hypothesis. The alternate hypotheses of an antecedent and a subsequent development of these water-gaps have also found champions, and a much closer examination is needed before a conclusion is arrived at.

A. W. Rogers (1925, p. 10) conveniently sums up the three alternative explanations that have been given for the behaviour of these rivers, thus: "(1) that they began to run before the rocks forming the mountains were folded and that they held their courses during the uprise of the mountains themselves; (2) that they began their history when the folded rocks were buried under a cover of unconformably overlying sediments and that their valleys were sufficiently established when the folded rocks became exposed to prevent the harder rocks turning the rivers along their strike; and (3) that they worked their way inland by headwater erosion from shorter rivers draining the original slope backing the shore."

The first explanation, that these rivers are antecedent to the Cape folding, is refuted by the relative ages of the folding and the initiation of the rivers. De Villiers (1944) shows that the folding was initiated during the deposition of the Karroo beds and was completed before the end of the Karroo cycle. Drainage at this time was probably northwards as this region appears to have been the provenance of the sediments of the Molteno beds. The north-south drainage is in all probability later, and has been shown to be at least post-Enon in the Oudtshoorn region (J. W. du Preez, 1944). This is further borne out by the complete absence, within the Enon Conglomerates, of any fragments of rocks, particularly Karroo dolerite pebbles, that are definitely derived from the country north of the Swartberge. Apparently, the original post-Karroo drainage of this region was towards the east and was responsible for the removal of the Karroo strata and the opening up of the parallel synclinal valleys between the anticlinoria of resistant Table Mountain Series. Only after the Enon sediments had been deposited in these valleys were the poorts opened by the north-

south drainage, and an origin of these rivers that relies either on antecedence to the Cape folding, or on superimposition from the Karroo strata, is impossible. Rogers (1925, p. 11), however, believes that superimposition may have occurred from "a nearly continuous sheet of Cretaceous rocks lying on the folded strata", but there is no evidence to point to the existence of such a cover.

The only alternative explanation for the origin of these poorts then seems to be that they were cut in post-Enon times by the headwater erosion of rivers working their way inland. There can be little doubt that those Cretaceous tributaries of the east-west drainage which rose on the newly formed fault scarps on the southern flanks of most of these ranges, were swift and vigorous streams and that their headwater erosion was greatly aided by the open jointing of the folded strata. By working their way along these master-joint directions, many accomplished a breaching of the resistant ranges and proceeded to capture much of the drainage to the north of the respective ranges. This would explain the precipitous nature of the walls of the poorts, their twisting courses and the absence of high-level erosion benches on their flanks. The latter, as well as more open poorts, would be expected, had their origin been due to either antecedence or slow superimposition. It is also possible to find many streams still in the process of poort-formation along these lines, and all stages of their development may be observed.

The only possible exception to this general rule could be the Touws-Gouritz River, whose straight, open poort through the Langeberg-Outeniqua mountain range shows little or no control by joint systems and has a more mature appearance when compared with the other poorts. A possible explanation for this is that the Touws River was an established river before the anticlinal crest of Table Mountain Sandstone, now forming the Outeniqua range, was uncovered and that its lower course, the Gouritz River, had at this early date flowed across the buried range. Later uplift could then have been responsible for the incision of this transgressive portion of the river across the axis of this fold with the formation of the wide, superimposed Gouritz Poort. This explanation is, however, wrought with difficulties, not the least of which is that the region to the south of the present coast-line, towards which the Touws River would have been draining, was then probably occupied by highlands. It appears more likely that the initiation and early development of the Touws River is analogous to that of the other members of the north-south drainage, except that the breaching of the Gouritz Poort took place earlier than the formation of the other poorts. The collecting basin of the sediments derived from its valley behind the mountain chain and transported out of this valley would then be the Cretaceous basin situated around Herbertsdale. The Gouritz Poort would then also be a subsequent poort, from which, because of its greater age, much of the original structural control has been removed.

It appears, therefore, that these supposed "antecedent" or "superimposed" poorts are, in actual fact, subsequently controlled by the directions of the major joint systems across the anticlines, and that the north-south rivers, which have been responsible for their formation, are, on the whole, of post-mid-Cretaceous age and have derived their importance by the capture of the older east-west drainage.

(*e*) *Superimposition from erosion surfaces and degree of superimposition.*

Two important points, which warrant further consideration and which can now be more satisfactorily discussed than before actual examples of superimposed rivers were examined, arise from the above description.

Firstly, it will be noticed that no mention has been made of examples of the superimposition of rivers from particular erosion surfaces, bevels or peneplanes,

although evidence of the existence of these ancient surfaces in Southern Africa is widespread. Such a type of superimposition has frequently been described, but this conception is fundamentally unsound since, by it, the mere incision of a river into its own flood-plain or erosion surface qualifies it for classification as a superimposed river. Reference to the definition of superimposition, both as set out by Johnson (1932, p. 493) and as adapted for this paper, confirms the view that superimposition can take place only from one distinct formation on to another showing a structure foreign to the first. In other words, for superimposition it is imperative that the river passes, downwards, through an unconformity separating two distinct structural units. Should a river, whose course was initiated or established on an ancient erosion surface, suffer incision into this surface as a result of uplift, and should this vertical migration of the river carry it through an unconformity, then only are the conditions for superimposition satisfied, and the development of the river may be described as superimposed from the ancient erosion surface. Need it be emphasised that a river, with a similar history but which has not passed through an unconformity, could, at the most, be described as having a course *inherited* from the ancient erosion surface?

The second point which must be considered is whether there is a limit to the amount of adaptation a superimposed river may show to the foreign structure and yet remain superimposed. It was noted, for instance, that the southerly migration of superimposed portions of the Vaal River is actually a form of adaptation of the river to the structure of the unconformity. Further reference to the definition, however, shows that "superimposition" is a term which is descriptive solely of the historical evolution of the river, and that a river is either superimposed or not. No limit is set on to the amount of adaptation displayed towards the foreign structure, except that the course of the river should retain its general shape and direction of flow. It is, however, felt that certain degrees of comparison should be introduced to distinguish between superimposed rivers showing adaptation to varying extents. Thus, a "distinctly superimposed" river shows little or no adaptation; a "clearly superimposed" river is one whose position and form has been changed in detail only; and a "poorly superimposed" river shows extensive adaptation. On this basis, the lower course of the Kafferkuils River and the portion of the Vaal River across the Vredefort Dome become distinctly superimposed; most of the rest of the Vaal River and the Olifants River Poort through the Drakensberg of the Eastern Transvaal show features of clear superimposition; while much of the rest of the primitive, interior drainage has been poorly superimposed. It is acknowledged that in any descriptive natural science it is, of course, impossible to draw hard and fast boundaries, but it is felt that judicious usage of this arbitrarily divided scale would greatly simplify accounts of superimposed rivers.

VIII ANTECEDENCE IN SOUTHERN AFRICA

From definition, a river is antecedent only if it flows across a foreign structure which is both younger than the river and tends to obstruct it. The converse is also true: any rivers, crossing against the trend of a certain, given structure, are antecedent towards the formation of that structure, provided that they are older than the structure. To examine antecedence, it is, therefore, first necessary to determine where such structures occur, and then to examine their effects on the older rivers crossing them. It is, naturally, unnecessary to examine structures which came into existence before the initiation of our present drainage.

Broadly speaking, post-Karroo structures, towards which South African rivers may show antecedence, are of two types: crustal disturbances which have tended to

disrupt the rivers; and the covering of the land surface by Kalahari Sands which has had a choking effect on the rivers. The former may be further subdivided as to the forces causing the disturbances, and on this basis we may distinguish between disturbances resulting from the forces of continental drift, or "Gondwanide diastrophism", and those caused by the adjustment of isostatic equilibrium, or "crustal warping".

(*a*) *Antecedence towards Gondwanide Diastrophism.*

The final disruption of the Gondwanaland continent occurred in the mid-Cretaceous period and the forces of separation were manifest by large-scale marginal faulting in directions roughly parallel to the present South African coast-line. This normal faulting was a product of tension and the downthrow was towards the outside of the land mass. The most extreme throw is registered in the south where the faulting is located along another tensioned zone, the southern limbs of the overfolded anticlinoria of the Cape Fold belt. Here, the phenomenal vertical movement along the fault planes produced a series of gentle, north-south cross-flextures on the downthrown blocks, dividing the intermontane, synclinal valleys into many independent and disconnected depositional basins for the material being derived by the denudation of the fault scarps and added to the earlier accumulations of Enon deposits.

Contemporaneous with the diastrophism, the continental drainage was mainly that of the primitive interior rivers, which, since they nowhere appear to cross the structures resulting from this diastrophism, do not warrant further consideration. In the southern Cape Fold-belt, the through north-south drainage was not yet in existence, it being likely that the very initiation of these rivers depended on the east-west faulting to supply the steep scarps on which they could take their rise and commence their breaching of the mountain chains. The Touws River is the only river that could possibly have crossed any of the mountain ranges at this early date, but whether it did so is still uncertain. Nor is it certain whether the Worcester Fault may be continued along the southern flank of the Langeberg-Outeniqua range to cut the lower reaches (the Gouritz River) after it emerges from its poort through these mountains. Should this be permissible, and, should the Gouritz River Poort be proved to be a superimposed poort, then the Touws-Gouritz River system must be considered antecedent to the Gondwanide faulting. The general existing drainage of the intermontane valleys was, however, the east-west drainage, and the very fact that the cross-folding divided the valleys into completely disconnected and independent Enon-depositional basins, shows that the disrupting influence of the cross-folding was sufficiently severe to completely dismember the existing drainage. The only examples of antecedence to these cross-flexures that have been described are displayed by the winding gorge of the Baviaans River at Beako's Nek, and the passage of the Breede River through the Goree Poort (M. S. Taljaard, 1949, pp. 72 and 144).

(*b*) *Antecedence towards Crustal Warping* (See A. L. du Toit, 1933.)

The removal, by erosion, of vast quantities of material from the land mass must necessarily have lightened it considerably, and the deposition of this waste on the adjacent sea-floor must have weighed it down. This disturbance of equilibrium would be balanced by an isostatic adjustment whereby the rising of the interior would accompany a deepening of the continguous oceanic basis. Updoming of the continent would further be greatest where the most material had been removed, and, conversely, heavy laden portions of the crust would not register upheaval to a comparable extent. Such adjustment of isostatic equilibrium took place during the late-Cretaceous and late-Tertiary periods, and, especially during the latter, the highly differential nature

of the uplift dimpled the interior into a series of depressions separated by long swells or ridges of uplift.

The greatest amount of erosion was suffered by the coastal fringe and uplift appears to have reached a maximum just behind the shore-line. It is likely, therefore, that a complete line of warping stretches around the margin of the continent just inland from the coastline. In a number of localities this warp-axis has been traced, and the rivers crossing this axis are almost always found to be antecedent to this warping. The example of the antecedence to the warping displayed by the Kafferkuils River has already been described, and many other rivers show similar, though usually inferior, development.

In the interior, the crustal deformation which accompanied the updoming gives rise to two distinct features: lines of maximum uplift or axes of uplift, and regions of relative depression or basins of depression. The latter usually occur where the crust has been heavily laden, such as about the Bushveld Complex in the Central Transvaal and the deepest portions of the Karroo geosyncline. To some extent the axes of uplift seem to occupy positions of pre-Karroo prominence, such as the Witwatersrand ridge, the Zoutpansberg highlands and the Khomas Highlands, but on the whole their north-east to south-west trend appears to have been determined by the forces which controlled the rifting of the Central African continent. The accompanying diagram shows the distribution of these crustal deformational features, and it is interesting to note how their pattern has determined the form of the individual interior drainage basins.

The axis of uplift, in particular, appear to have had a pronounced dismembering effect on drainage lines and at present they all appear to form important continental watersheds along almost their entire lengths. In fact, at only two points are rivers found which have not been dissected by them, and, at these points, the rivers' courses are naturally antecedent to the warping. These transgressive courses belong to the Orange River and its tributary, the Hartebees River, and both show antecedence to the Griqualand-Transvaal axis of uplift. As quoted by Wellington (1933, p. 63), the gradients of the Orange River both above and below Buchuberg, the point where it crosses the warp axis, show instructively the effect of this uplift on the floor of the river. Thus, behind the axis, between Prieska and Buchuberg, the gradient is very low, averaging about 1·9 feet/mile. On crossing the axis at Buchuberg, a steepening of the gradient to 8·5 feet/mile is experienced, after which a value of about 3 feet/mile is maintained as far as Upington. Undoubtedly, the ridging up about Buchuberg has been responsible for the flattening experienced by the river channel before the axis is reached. This feature may be even better observed on the Hartebees River where the very low gradient, above the intersection with the axis at Kenhardt, has resulted in a temporary ponding and the formation of vast clay-flats such as Groot Vloer and Verneuk Pan.

Antecedence towards the basins of depression is, on the other hand, a fairly wide-spread feature, and is especially well shown by the rivers crossing the Bushveld Basin of the Central Transvaal. By the middle of the Cretaceous period, stripping of the Karroo strata had exposed prominences of the pre-Karroo landscape in this region, and it is probable that, in the south, the Witwatersrand already formed an important watershed between the Olifants, the Limpopo and the Molopo Rivers to the north, and the Vaal River to the south. River gravels of this age show that the Central Transvaal was shared by the hydrological basins of the Limpopo and the Molopo Rivers, and this drainage to the north and to the west flowed over a normal slightly convex surface. Since that time, however, this central portion, weighed down by the relatively heavy igneous rocks of the Bushveld Complex which form this part

of the earth's crust, has undergone very gradual, but intermittent, sagging. The subsidence of the floor of this depression, amounting to as much as 1,000 feet in places, shows some inequalities in sinking by the presence beneath the Springbok Flats of at least two subsidiary warpings, whose axes are, like those of the main basin, directed east-north-east. A further effect of this sagging must have been to increase the centro-clinal dip of the resistant strata forming the rim of the depression, thus rendering the continued flow of the rivers still more difficult. To maintain effective drainage across and out of this sunken depression, the rivers had to remain active and many of the more feebly established rivers were disrupted and dismembered, their headwaters joining those of the larger rivers. By actively incising their courses in the regions to the north of the depression, and often by building up their stream-beds across the flats by alluviation, the more firmly established rivers, such as the Elands, the Olifants, the Crocodile and the Nyl Rivers, have retained their northerly flows across the depression. Especially in their passage through the encircling highlands, downcutting has been sufficiently vigorous to keep pace with the sagging of their stream-beds higher upstream. These rivers are, therefore, antecedent to the crustal sagging which formed the Bushveld basin of depression.

Similar, but not nearly as pronounced, antecedence is shown by the upper Orange River and many of the southern tributaries of the middle Orange and the Vaal Rivers where they flow across and out of the Karroo-Basutoland structural basin of depression. The Zambezi River shows clear antecedence to the depression of the Okavango-Linyanti-Ngami basin, which it crosses between Katima Molilo and Katambora. Here the river countered the gradual sagging of the crust across its course by the aggradation of a wide flood-plain. Slightly to the south, however, the sinking of the Makarikari depression combined with the arching up of the crust along the Rhodesia-Kalahari axis to break the flow of the Okavango River into the Limpopo River, ponding it back into the lake occupying much of the Makarikari Depression.

(*c*) *Antecedence towards the spreading of the Kalahari Sands.*

It was shown on pages 13 and 14 that, during late-Tertiary times, the spreading of a cover of Kalahari Sands over the drainage lines of the interior rivers resulted in a foreign structure which tended to interrupt the continued flow of these rivers by choking them. It was further explained that the larger rivers, experiencing periodical floods, were able to keep their channels clear and retain their original flow-lines, showing thereby antecedence towards the covering by these Sands. Such a history is indicated for many of the interior rivers, most notably the upper Zambezi, the Okavango, the Molopo and the Nossob Rivers, and it is also felt that many of the tributaries of these rivers may show similar antecedent development, although the possibility of their being superimposed consequent from the surface of the Kalahari Sands must not be overlooked.

IX COMBINED SUPERIMPOSITION AND ANTECEDENCE

The breaching of the highlands encircling the Bushveld Basin of the Central Transvaal has been variously ascribed to the superimposition of the northerly and westerly flowing rivers, and to their antecedence to the sagging of the crust along their upper courses. These highlands are composed of the resistant quartzites of the Transvaal System, which possess an inwardly-directed dip. Prior to their burial beneath horizontal strata of Karroo age, these rocks must already have possessed a centro-

clinal attitude, although their dip was not as steep as it is today, and, as foreign structures, they therefore antedate the initiation of the rivers—an essential requirement for superimposition. However, the extent of their present intersection as well as their prominence as obstructing foreign structures, may be ascribed to their situation along the hinge-line of the depression. Since some of the warping has occurred after the initiation of the present drainage, the drainage must necessarily be antecedent to the warping. Therefore, both superimposition and antecedence have combined to produce the breaching of the highlands surrounding the Bushveld Basin by the Crocodile, Steelpoort and Olifants Rivers. In the case of any one river, the amount of superimposition may be taken as proportional to the tilt possessed by the underlying strata just before they were exposed, while the amount of antecedence would be represented by the extra inclination given to the beds after the initiation of the particular river.

For combined superimposition and antecedence, there must, therefore, be positional coincidence between the buried foreign structure and the later structure, and it is doubtful whether these conditions have been fulfilled elsewhere in South Africa.

X CONCLUSION

The most notable characteristic of superimposed and antecedent rivers is, naturally, that feature which serves to distinguish them from the normal rivers: viz., their transection of structures which, normally, would tend to give rise to very different types of drainage. But we have seen that this feature is by no means essentially developed in superimposed rivers, in which it is only the coincidence of the river being let down on to such a structure that gives rise to such marked abnormality. Likewise under special conditions, ordinary subsequent streams may, by normal erosion processes, develop courses which show an apparently similar, transgressive attitude to extraneous structures. It has been the failure to recognise these two facts which, in the past, has resulted in so much confusion in the description of river development.

In Southern Africa, superimposition of rivers has generally occurred only from two different stratigraphical formations—rocks of the Karroo System and the raised marine deposits of post-Karroo age. Superimposition from the former is mainly confined to the primitive, consequent, interior drainage of the sub-continent, while only the extended-consequent lower reaches of the rivers may display superimposition from the raised marine deposits fringing the continent. Just as the ancient, foreign structures, on to which superimposition has occurred, vary greatly as regards age, attitude, and lithology, so also do the rivers differ in competence and capacity, from time to time and from place to place. With all this diversity, it is not to be expected that the foreign structures will show any effect on the gradients of the rivers which could be considered as characteristic, but the slopes of the stream beds probably undergo the normal evolution that would be displayed by any incised, rejuvenated river in a similar structural environment.

Structures towards which South African rivers show antecedence are mainly confined to those produced by the crustal warping of post-Karroo age. On the whole, the positions of the axes of uplift have profoundly controlled the ultimate shape of the interior drainage basins and, except for the marginal warp axis running parallel to the coast-line, examples of rivers antecedent to them are scarce. Where such axes are crossed, the profile of the stream-beds shows an upward-convexity, which, as du Toit has pointed out, has deprived South Africa of large navigable rivers and wide-

spread arable lands. On the other hand, antecedence towards the basins of depression is fairly common, and the antecedent courses across these depressed areas are usually built up on alluvium and river-flow is sluggish. The rims of these basins, across which the antecedent rivers leave these depressions, produce effects on the rivers' gradients which are comparable to those produced by axes of uplift.

Unfortunately, while age relations remain vague, finality on the opposing theories of a superimposed or an antecedent development for the rivers which have been affected by the spreading of the cover of Kalahari Sands cannot yet be reached.

Finally, although crustal warping in the late-Cretaceous and late-Tertiary times has to a certain extent controlled the placement of most of the interior watersheds, it is due to the processes of superimposition and antecedence that much of our primitive drainage has been preserved, and that we are thus more easily able to examine the post-Karroo evolution of the landscape of the sub-continent.

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THE DIORITES OF YZERFONTEIN, DARLING, CAPE PROVINCE

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(Submitted November, 1951)

ABSTRACT

The recent suggestion of the existence of a basaltic magma in the South-west Cape Province shortly before the intrusion of the Cape granites is corroborated by the occurrence of a gabbroic body of pre-granite age at Yzerfontein. A detailed study of the primary rhythmic banding and igneous lamination of this body indicates its original form as being sheet-like or laccolithic, and the differentiation appears to have been due to a combination of fractional crystallisation and gravitative crystal settling on to a sub-horizontal floor. The crystal settling process was probably later arrested by the viscosity of the rest magma, due to its enrichment in potash and alumina.

The Yzerfontein diorites proper represent a hybrid product resulting from the mixing, in depth, of the marginal contaminated facies of the Darling granite with gabbroic material. The present uniformity of the diorite is ascribed to the emplacement of the hybrid magma to a higher level.

Gabbroic xenocrysts are widely distributed in the diorites, these single crystals almost invariably being surrounded by rims of material later in the reaction series. Amongst the mafic minerals a uniform orientation relation is found to exist between the cores and rims of such reaction products.

Two different types of granite aplite and pegmatite, namely potash-rich and soda-rich varieties, have invaded both the diorites and gabbros along joints corresponding in direction to the joint system of the Darling granite pluton, while swarms of veins and dykes containing late hydrothermal minerals, which probably represent the final stages of the magmatic history of the granites, follow similar directions.

This paper was awarded the Corstorphine Medal and the first prize of the Geological Society of South Africa for the year 1951

CONTENTS

I. INTRODUCTION

A suite of basic, igneous rocks is found along the Atlantic coast-line of the South-western Cape Province in the vicinity of Yzerfontein Point. Although the existence of these interesting rocks has long been known, little more than passing reference to their presence has previously been made.

In his first report to the Geological Commission of the Cape of Good Hope, A. W. Rogers (1896), while dealing with the south-western dolerite dykes, mentions that "at Yzerfontein . . . there is a large mass of doleritic rock which differs in important respects from the other basic rocks of this district". In a later publication (Rogers, 1905) the same author describes "a large mass of hornblendic rock, coarsely crystalline, with a banded structure; some thick layers are formed entirely of green hornblende, and others, usually thinner, have a fair proportion of plagioclase in them" and their dioritic character is indicated. Although the rocky coast-line around Yzerfontein was correctly mapped as "diorite" on Cape Sheet 4 of the following year (1906), these rocks were not differentiated from the "younger Cape granites" on the later 1 : 1,000,000 geological map published by the Department of Mines (1925).

In his report on the phosphates of Saldanha Bay, A. L. du Toit (1917) mentions that a sample of limestone, collected in this vicinity by Dr. Rogers, carried a fair proportion of phosphoric oxide, while W. Wybergh (1920) in a subsequent investigation, could find no such rich phosphate deposits. Neither of these reports mentions the nature of the underlying rocks.

"Cliffs of diorite, 40 feet high" are described as breaking the long, sandy beach of Saldanha by A. V. Krige (1927) in an examination of the physiographic characteristics of the South African coast, and in the same paper attention is directed to the raised beach deposits situated to the south-east of the Point.

The most recent reference to these rocks has been made by D. L. Scholtz (1946) who mapped them as a distinct and separate rock-type, and suggested their hybrid relation to the younger Pre-Cambrian Granites of the Cape Province. He further showed that the pyritic veins and dykes which traverse these hybrid rocks, belong to an extension of the north-west trending zone of mineralisation stretching from Helderberg through Kuils River and the Koeberg Hills, and concluded that these ore deposits are genetically related to the younger granites.

Two similar occurrences of basic rocks are known in the South-western Cape Province; The Brewelskloof Diorites and the Malmesbury Diorites. The latter comprise four composite stocks, intrusive into a dome structure in the Pre-Cape sediments along the western side of the Malmesbury-Paardeberg granite pluton, and have recently been described by P. J. van Zyl (1950). Although intimately associated with the Cape Granites, these diorites are believed to be older than the granite intrusions and are found to have had a pronounced contact metamorphic effect on the sedimentary country rocks. Petrologically the rocks vary from normal gabbros to quartz diorites and are shown to owe their origin to successive intrusions of a pre-Granite basaltic magma which had undergone different degrees of differentiation in depth.

The Brewelskloof diorites, now being investigated by P. J. Joubert, also invade the Pre-Cape sediments and occur about 6 miles north-east of Worcester. Their com-

posite stock-like form may be indicated by a central, more basic portion surrounded by a more acid margin. The rocks have, however, been severely altered, presumably by regional metamorphism, no primary minerals remain and their original nature is therefore doubtful.

II. TOPOGRAPHY

Yzerfontein Point forms a small promontory protruding into the Atlantic Ocean approximately midway between Table Bay in the south and St. Helena Bay in the north. The nearest town, Darling, lies about 14 miles due east, while the important guano-collecting station, Dassen Island, is situated approximately 7 miles to the south-west. Along a narrow strip of the coast-line in the vicinity of the Point and for a short distance to both the north and the south are exposed rocks belonging to a complex igneous mass and varying in composition from basic olivine-gabbros, through more acid augite-diorites to aplites and pegmatites of granitic affinity.

Bold cliffs, often at least 40 feet high, have been cut into these rocks where they abut against the wild open sea. In addition, three distinct terraces, stepping the surface between the sea and the interior, are prominently developed. The lowest is definitely wave-cut and varies in height from 20 to 25 feet. This surface is most clearly seen cutting across "Die Eiland", a small peninsula to the south of the Point which is separated from the mainland by a narrow tidal channel. It may also be observed bevelling the four rocky outcrops along the beach stretching from Die Eiland towards the south-east, and it is on this level that the fishing factory at the Point and the holiday resort of Yzerfontein have been built. Comparable in height is the sandy terrace extending parallel to the coast for about a mile southwards from Die Eiland. Its width averages 100 yards and it is at least 26 feet high. This feature was noted by Krige (1927) who came to the conclusion that it "represents a true raised beach, as its surface is too uniformly flat to have been built up by storm waves during a considerable progression of the shore".

The second terrace varying from 40 to 60 feet above sea level is less prominent than the lower one, though comparable in area. It may best be seen cut in the dioritic cliffs to the south of the Point and it generally possesses a slight dip towards the sea. It is, however, by no means confined to this stretch, although its feeble development both towards the north and the south may be difficult to follow.

Rising fairly abruptly from this level, the land surface finally evens off to form the third and most distinct of the three terraces. This platform is remarkably flat and featureless, but possesses a very slight dip (less than 1°) towards the north-east. In this direction it attains its widest development and where it overlooks the resort it extends for approximately a mile inland with a constant height of 120 feet. From here it is prolonged in a south-easterly direction for almost two miles parallel to the coast-line, but before reaching its southern extremity its width narrows down to about 300 yards while its altitude increases almost imperceptibly to over 160 feet. (Plate I, No. 1).

All three platforms are blanketed by a veneer of sandy limestone or a covering of pale-coloured shelly sand, which conceals to a large extent the underlying igneous rocks. Frequently this unconformable cover displays basal lenses of dioritic pebbles and rounded boulders showing evidence of wear by wave action, and there is little doubt that they can be correlated with the raised marine deposits extending round the rest of the continent.

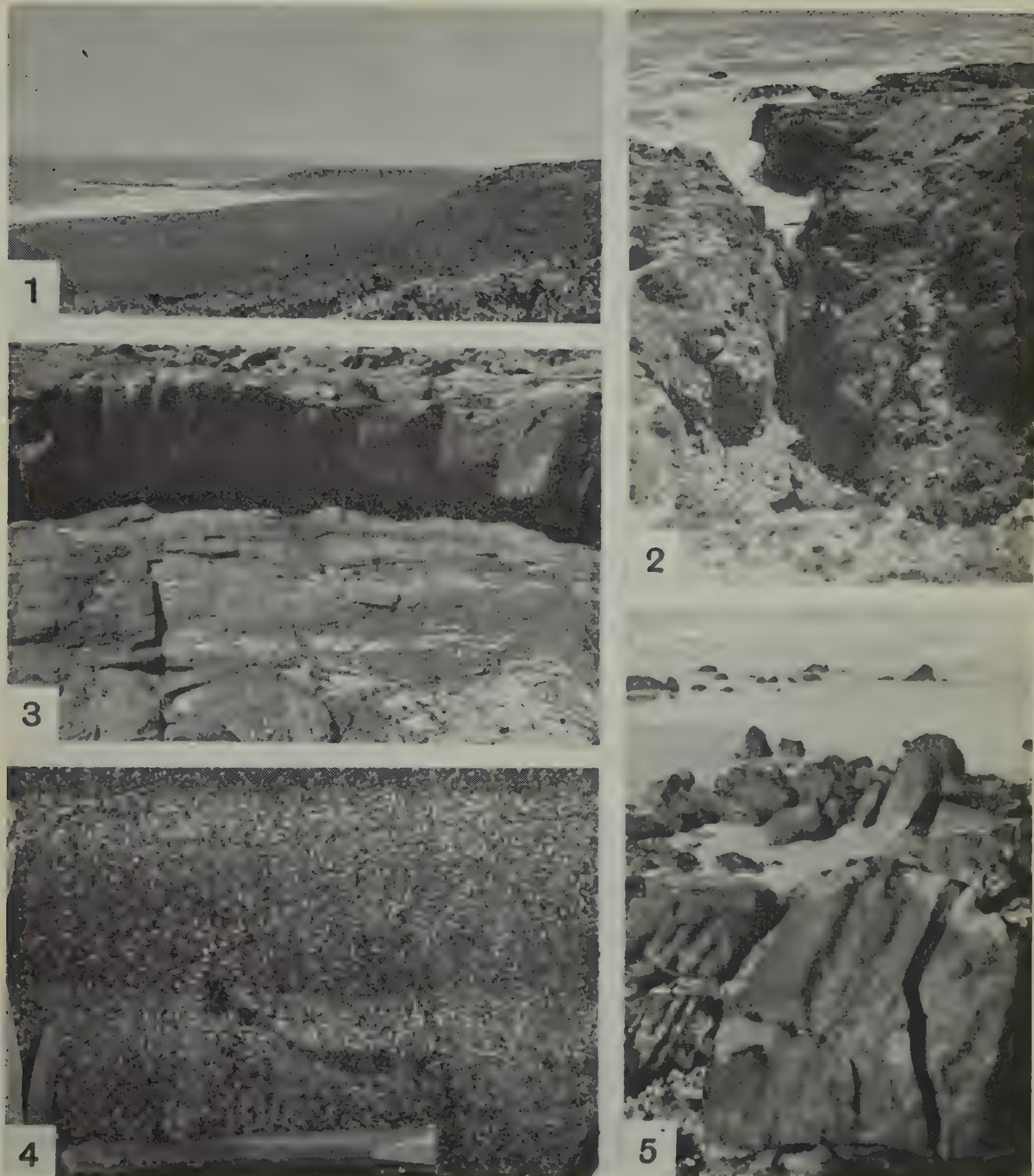


Plate I.

1. View of Die Eiland from the south. The 120-foot platform is to be seen on the right, while the raised beach deposits occupy the foreground.
2. Duiwe Nes—a deep, narrow gulley, situated along a thick pyrite-tourmaline vein.
3. Swarm of parallel, closely-spaced, tourmalinised veins. Later, flat-lying joints visible in the background.
4. Banded gabbro, illustrating clearly the density gradient across the individual bands. The print has been cropped so as to demonstrate the original, horiziontal attitude of the layering.
5. Banded gabbro outcrops at Gabbro Point showing the present, steeply-inclined attitude of the mineral banding.

The most prominent topographical feature in the vicinity is the conical hill, Vlaekop, which rises fairly abruptly from the 120-foot platform to a height of 280 feet. The summit of this low hill has been stripped of its concealing blanket of limestone and rocks belonging to the igneous complex are found surrounding the beacon which marks its crest.

Separating the high ground around Yzerfontein from the low granite hills of Slangkop and Koffiekop to the east, which are the nearest known exposure of the rocks of the Darling Pluton, is a 7-mile broad belt of level, sand-covered country, the elevation of which seldom exceeds 100 feet and averages less than 50 feet.

Although several traverses were made across this gap, not a single outcrop of rock was found and only occasionally were small rounded granitic pebbles encountered embedded in these sands. This sandy stretch continues towards the north-west and the south-east, and the nearest rock outcrops in these directions are respectively the granites at Geelbek and the rugged cliffs of Malmesbury rock at Bok Point. It will be shown later that most of this region, particularly the stretch between Vlaekop and Koffiekop, is probably underlain by the Younger Pre-Cambrian Granite of the Darling pluton, while the rocks of the dioritic complex are confined to the high ground around Yzerfontein and for a short distance to the north. Further, it appears that this high ground is built solely of diorite and its thin, recent limestone cover and there can be little doubt that the existence of this promontory, which is the only relief from the monotony of an otherwise smooth expanse of beach and sandy, often dune-covered, coastal plain extending for fully 25 miles from Saldanha Bay to Bok Point, is due to the superior resistance of the dioritic rocks to the forces of marine denudation when compared with the granite.

With the exception then of the summit of Vlaekop, rocks belonging to the basic Yzerfontein Complex are exposed only along the wave-washed coast in the vicinity of Yzerfontein itself. The most continuous exposure is that which forms the Point and extends in an almost uninterrupted sequence to the holiday resort on the one hand and to Die Eiland on the other, a distance of almost 3,000 yards, but the outcrop is nowhere more than 200 yards wide. Along the stretch of shelly beach further down the coast, four smaller exposures are found all within the first two miles from Die Eiland, while on the northern side of the resort a clump of rounded igneous boulders, 1,000 yards from the bathing shelter on the beach, form both an important and the only occurrence of these rocks in this direction.

III. THE GEOLOGY OF THE IGNEOUS ROCKS

Great variation exists in the grain size of the dioritic rocks. Generally the most basic varieties possess the coarsest grain with the average size of the crystals varying between 2·00 and 5·00 mm., while the colour index of these types in no way reflects the basicity of the rocks. This is due to the frequent development of primary mineral banding in these types. A much more uniform texture and homogeneous colour is displayed by the less basic varieties and these types possess a finer and more even grain. They are free from primary banding and do not display sharp intrusive contacts towards the more basic types. The change between the two types is, however, accomplished across a broad, transitional zone, but the frequent occurrence in the finer-grained type, near this transitional zone, of large, coarse-grained, ghostlike inclusions with either a lighter or a darker colour may be taken to reflect their younger age relative to the more basic types. The most acid rock-type found is a very fine-grained, pink,

aplitic dyke-rock which always shows sharp intrusive contacts towards the darker. more basic varieties. Occasionally too, coarse-grained pegmatites, comparable in all respects with the aplites, are developed.

On the basis of the above variations, as well as on mineralogical grounds, it is possible to recognise the following suites of rock types at Yzerfontein:—

(i) A coarse-grained gabbroic suite.
(ii) A suite of heterogeneous, transitional rocks.
(iii) A fine-grained dioritic suite.
(iv) A suite of later aplitic and pegmatitic injections.

Phase petrology has played an important part in this distinction of the various rock types, but the variation of the phasal compositions of the different suites will be discussed only after their distribution and petrological characteristics have been described.

(i) The Gabbroic Suite.

Although it is believed that the gabbroic rocks must originally have constituted a fairly large body, the present outcrops of these rocks are very limited. Only at the clump of boulders on the beach to the north of the resort do they occur in any quantity. but similar gabbroic types are also found on the northern edge of the lone outcrop near the bathing shelter, on the seaward extremity of some of the rocks jutting into the sea below the resort, at the "Poikilitic Gabbro Promontory" and on Meeuw Rock. the bevelled, rocky, guano-covered island in the middle of the bay.

The rocks of this suite are characteristically coarse-grained gabbros, frequently banded and composed essentially of well-formed, tabular crystals of pyroxene and plagioclase, with flakes of biotite and interstitial orthoclase also visible in hand specimen.

The pyroxene grains, however, are very dark, almost black, particularly in slightly weathered specimens, and their general appearance in the field is more suggestive of hornblende than pyroxene with the result that these rocks have been erroneously referred to by Rogers (1905) as diorites.

Primary rhythmic banding is not universally developed in these rocks and at only two localities could the orientation of this structure be measured. Here the banding is seen to be due to the varying proportions of the essential minerals. (Plate I, No. 4.) Thus types vary from a dark, almost black rock, extremely rich in pyroxene, to a leucogabbro in which the mafic minerals are very subordinate to the feldspar. The thickness of the individual bands is remarkably constant and averages nine inches in width, while the boundaries between the bands are sharp and distinct. A density gradation is, however, developed across each band, whereby the lower side with reference to the dip of the banding is always dark and pyroxene-rich, while the lighter, felsic minerals appear to be concentrated in the upper portions of the layers. Also apparent in the field is the arrangement of the large crystals of plagioclase and pyroxene in such a way that their greatest dimensions parallel the plane of banding. This feature is described as "igneous lamination" by Wager and Deer (1939) and seems universally associated with primary rhythmic banding.

Occasionally too, as at the "Poikilitic Gabbro Promontory" and on Meeuw Rock, a distinct facies of the non-banded gabbro is developed which, when viewed from a distance, presents a peculiar mottled appearance. (Plate II, No. 1.) On closer inspection the dark spots responsible for this feature are found to be large, rectangular insets of orthoclase up to 2 inches long and $1\frac{1}{2}$ inches wide, and possessing good crystal outlines. These insets are, however, not homogeneous but are liberally studded with

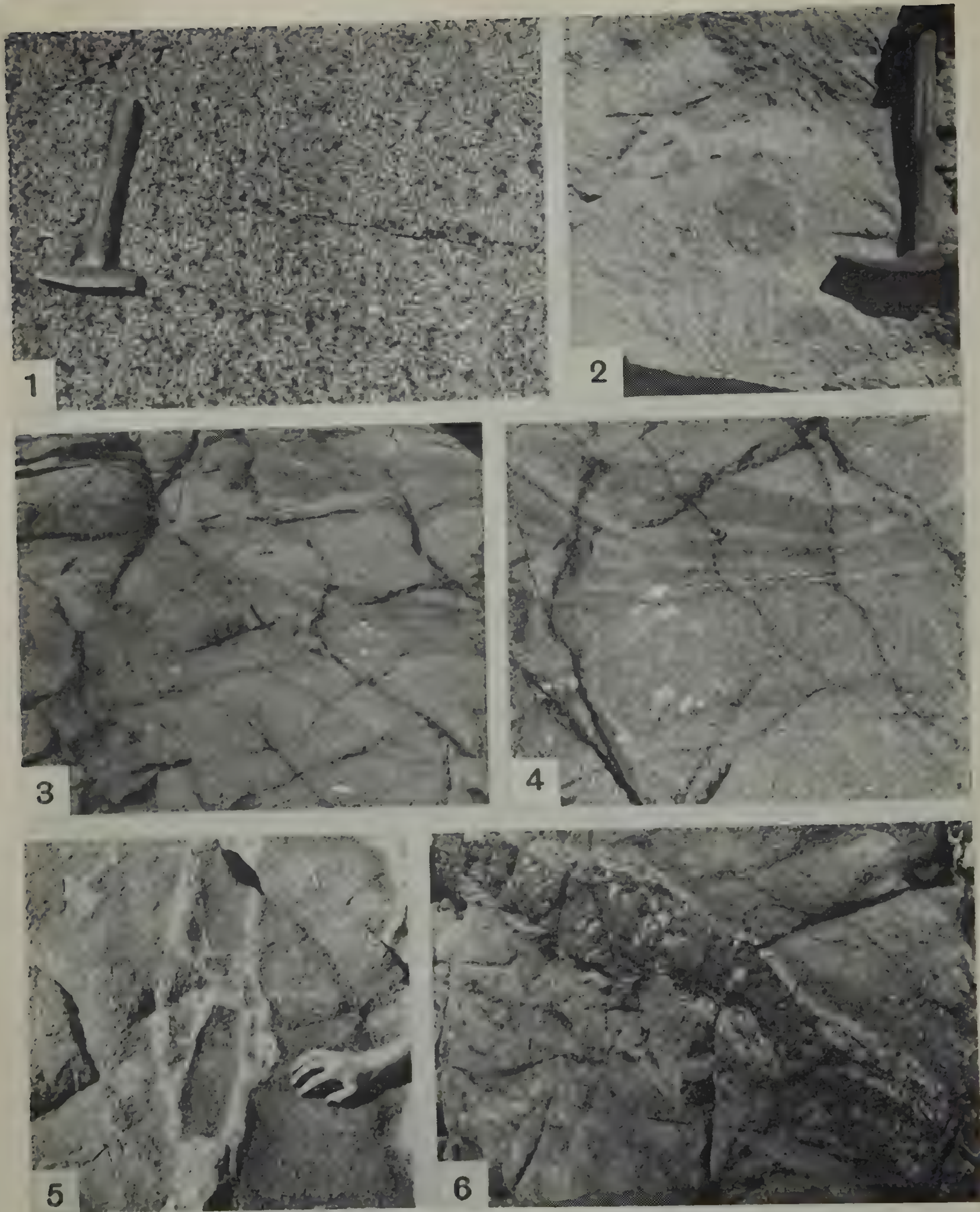


PLATE II.

1. Mottled appearance of the poikilitic monzo-gabbro.
2. Rounded inclusions of the gabbro chill phase in diorite.
3. and 4. Mixed aspect of the transitional zone.
5. Aplite dyke containing scattered fragments of diorite.
6. Tourmalinised shear-zone.

inclusions of well-formed crystals of pyroxene, plagioclase, biotite and iron ore. On either side of the Promontory and on the south-western edge of Meeuw Rock, this facies grades imperceptibly into more normal gabbro by a decrease in the amount and size of the poikilitic orthoclase crystals, and this, coupled with the similar mineral assemblage and chemical composition, justifies our regarding this "poikilitic gabbro" not as a separate rock type but as a facies of the normal gabbro suite.

(ii) **The Transitional Suite.**

This zone occupies a belt situated between the gabbros and the diorites, the outcrop width of which varies from 10 feet in the north to at least 60 yards in the south. Two main occurrences are observed: the northern occurrence stretching from just below the offices of the Atlantic Ocean Industries for approximately 550 yards towards the north-east where it disappears beneath the sands just below the resort; and the western occurrence which may be traced along the westerly extremities of many of the promontories between Duiwe Nes and Die Eiland, a distance of 1,200 yards. As the rocks of these two belts are identical, they probably extend out to sea to join somewhere to the west of the Point around which they curve.

The rocks of the transitional suite may truly be referred to as "mixed rocks". In the field they have a patchy and uneven appearance and much variation from place to place is seen in the proportions of the constituent minerals, the colour and the grain size. Generally the coarse-grained types appear to belong to the gabbroic suite, while the finer-grained more homogeneous varieties cannot be distinguished from the dioritic rocks in hand specimen. These various types are intimately mixed. Sometimes large rounded ghostlike inclusions of the coarse gabbroic rocks are found in the finer-grained type, while at other times there is an intricately streaked and whorled relation between the types which may be likened to the pattern obtained on partially stirring up a mixture of thick, coloured paints. (Plate II, Nos. 3 and 4.) Although both gabbroic and dioritic types are distinguishable in these streaks and whorls, even here sharp contact between the types is absent and transition from one to the other is gradual.

The rocks of this suite are intermediate in character, both chemically and mineralogically, between the gabbros and the diorites.

(iii) **The Dioritic Suite.**

Practically all the remaining outcrops at Yzerfontein are formed by rocks of this suite. They are readily distinguished by their greater homogeneity, constant light grayish green colour, fine even grain (1-3 mm.) and total lack of any primary structures. Feldspar is by far the most abundant mineral in these rocks. The pyriboles appear in the form of relatively large, green, sugary grains which may occasionally reach a length of 5 mm. Less abundant are flakes of mica which have often been altered to green chlorite and are easily confused with the other dark minerals. Quartz is not visible in hand specimen.

The rocks of this suite are apparently the most resistant of the complex and the configuration of the coast reflects to some extent the trend of their junction with the transitional and gabbroic suites. The outcrops are prominently developed around the Point and extend south-eastwards to include the four smaller headlands lower down the coast. In addition, outcrops are found at various points below the resort while the rocks are well exposed around the Vlaekop beacon.

Although characteristically homogeneous, the dioritic rocks often contain inclusions of the older gabbroic suite. These are mainly confined to the coarse-grained

ghost inclusions near the transitional zone, but at some localities a diorite is developed which is extremely rich in small, rounded inclusions averaging 4 inches in diameter. (Plate II, No. 2.) This "inclusion-bearing diorite" is well displayed in the quarries at the harbour where the inclusions are composed of a very fine-grained rock, slightly darker in colour than the surrounding diorite towards which they show sharp boundaries. On chemical and mineralogical grounds, these inclusions belong to the gabbroic suite. Their anomalous grain size and xenolithic development will be discussed later. Prolonged search of the quarries brought to light only two inclusions of coarser grain and darker colour which proved to be the more normal members of the gabbro suite.

(iv) **Aplite and Pegmatite injections.**

The aplite consists of a distinct pink rock with so fine a grain that the individual crystals cannot be distinguished with a hand lens. Occasionally it is possible to pick out extra large clusters of tourmaline, dark green hornblende or chlorite, but even these have sizes usually less than 0·5 mm.

The aplite injections are found to fill veins, fissures, dykes and irregular pockets in the rocks of all three other suites. Towards these rocks they show sharp contacts and they have also altered them for a distince of about three inches from these contacts, manifest either by a pink staining or by the whiter colour shown by hydrated feldspars. Injection has often been accomplished by the shattering of the intruded rock and the aplitic material may engulf angular chips and fragments up to 6 inches long of the adjacent rock. (Plate II, No. 5.) Sometimes these fragments show edges which may be pieced together after the style of a jig-saw puzzle. Although injections of this type are highly irregular along their length, some aplite dykes are found which have produced little shattering and have a more constant width and direction over long distances. They have exploited the two prominent sets of joints striking approximately 138°-144° M.N. and 6°-12° M.N. Still another mode of occurrence is displayed on the western side of the region where, roughly midway between the Point and Die Eiland, aplitic material has been injected into the diorite, not as dykes or veins but as a large irregular mass from which may radiate dykes and veins. Similar, though very much smaller, isolated aplite bodies are frequently encountered on Die Eiland and these aplite-filled "pockets" often carry large rounded foreign inclusions ranging up to 9 inches in diameter. Although mainly dioritic, many of these inclusions are of coarse-grained gabbro while an occasional hornfelsic type is clearly composed of indurated argillaceous sediments. This must show the proximity of both the gabbro body and the sedimentary rock into which it was intruded, although outcrops of these are here missing.

A pegmatitic facies was discovered by Prof. Scholtz who kindly gave the author a large sample of this coarse-grained pink rock. The rock is predominantly feldspathic, crystals of this mineral usually having dimensions of 2×1 cm. Much quartz occurs scattered interstitially and with it are associated blebs and spangles of pyritic ore. Patches of the rock are greatly enriched in epidote and around these green portions feldspar alteration has been extreme. A similar rock type is frequently found elsewhere but usually only in the form of narrow irregular stringers and veins. Those towards the north of the quarry do not usually show the pink colouration, while towards the south epidotisation becomes pronounced and colour is more strongly displayed. Also towards the south, injection has been accomplished by much shattering of the adjacent rock, fragments of which are included in the pegmatite veins.

Late Hydrothermal Injections.

One of the most striking features observed in the rocks to the south of the Point is the large number of veins which have been filled with late hydrothermal minerals. (Plate I, No. 3.) These veins are everywhere present and are found cutting all the various rock types. Along the coast between the Point and Die Eiland where these veins have been most prominently developed, the spacing between the individual veins is seldom more than a foot, while their thickness varies from a few millimetres to well over a foot. To the north the veins are generally more widely separated and the hydrothermal products are present as thin veneers along joint planes, seldom being visible without the aid of a lens. Towards the south from Die Eiland a similar decline in their development is apparent.

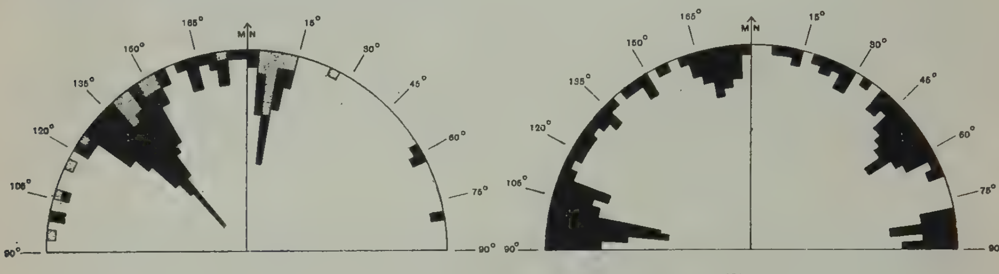
The minerals occurring in these veins include tourmaline, pyrite, jasperoid, quartz, muscovite, calcite and epidote. With the exception of quartz, these minerals are never found to occur singly in the veins. Thus calcite is usually accompanied by epidote, muscovite by quartz, pyrite by tourmaline, quartz or jasperoid. In the larger veins these associations take on the form of rhythmic crustification banding, while in the thinner veins a random distribution of the minerals is observed. An indication of the relative ages of the minerals is given by this banding and by the occasional intersection of the veins, which shows that the tourmaline and pyrite were introduced before the jasperoid and that the epidote and calcite marked the final hydrothermal stage.

The introduction of this late hydrothermal material was by means of aqueous solutions and this has invariably resulted in the alteration of the adjacent rock. In the field this is manifest either by the chalky white appearance of the feldspars or by the pink staining of this mineral, while under the microscope the cloudy, saussuritised feldspars may show the presence of fine, dusty hematite and are frequently in the process of being replaced by tourmaline, epidote and calcite. Extensive chloritisation of the dark minerals is also a feature of this alteration. Perhaps the difficulty experienced in finding suitable, unaltered specimens of diorite south of the Point may be attributed to the extensive alteration of these rocks by the closely spaced veins in this region. Either the presence of the hydrothermal material in the veins or the alteration of the adjacent rock seems to have lowered the resistance of these directions to the attack of the elements, for the numerous steep-sided gullies on the western coast are found to parallel the general directions of these veins and invariably such veins of greater than average thickness are found at the heads of such inlets. (Plate I, No. 2.) The broad tidal channel separating Die Eiland from the mainland is also developed along two exceptionally thick pyrite-tourmaline-jasperoid veins.

There is little doubt that the material filling these veins is connected with the late hydrothermal mineralisation along the zone extending from Helderberg through Koeberg to Yzerfontein and which Scholtz (1946, p. lxxviii) suggests is related to the intrusion of the Cape granites and represents the products of a single metallogenic epoch. In this connection, Scholtz records the presence of 4 penny-weight of gold and subordinate chalcopyrite in some of the massive pyritic quartz-tourmaline veins. A study of polished sections of these veins from the quarry, however, failed to reveal the presence of these ores, but frequent green malachite staining of some of the veins towards the south indicates the existence of copper compounds. It is interesting to note that chalcopyrite and blende have been observed in the granites of the Schaapenberg quarry. (Scholtz—personal communication.) Tin mineralisation evidently did not occur as far north as Yzerfontein and panned concentrates revealed no cassiterite.

STRUCTURE.

The directions over 350 joints and veins have been measured and their orientations are shown by Figs. 1 (*a*) and (*b*).



(*a*) (*b*)

FIG. 1.

Plot of the strike directions of 350 joints.

(*a*) Stippled—aplite dykes, solid black—veins bearing hydrothermal minerals.

(*b*) Barren joints.

From Fig. 1 (*a*), which shows the directions of those joints that have been filled with vein-material, either aplitic or hydrothermal, two systems are clearly defined. The older with a strike varying between 3° and 15° is the most important aplite direction, while the younger which is best developed between 138° and 141° is the most prominent direction south of the Point. Although aplites do sometimes follow them, these latter joints have been chiefly exploited by the later hydrothermal veins, and, without exception, all the larger veins follow this set.

Fig. 1 (*b*) represents the orientation of the younger joints along which no late, hydrothermal minerals were observed in the field. Here four definite sets are apparent, the oldest striking at approximately 170°. It is quite possible that the joints of this set are also filled with thin films of the hydrothermal material which may easily have been overlooked in the field. The 170°-set could then be introduced into Fig. 1 (*a*) where it would be intermediate between the 140° and the 9° systems. The orientation of interesting crush-breccias parallel to this direction is noteworthy. These breccias assume the form of broad veins filled with angular fragments of diorite and fine, black mylonite, and have been extensively tourmalinised. (Plate II, No. 6.) Both in appearance and alignment they parallel similar features observed at many localities in the Western Province granites, particularly near Cape Columbine.

The younger joints of the remaining sets, striking at approximately 56°, 82° and 97° are less persistent in length than the earlier joints, which they have often displaced. Finally, a further, nearly horizontal joint-direction, younger than all other sets, is ubiquitous. (Plate I, No. 3.)

Unfortunately, no flow structures in the diorites could be detected either in the field or in orientated thin sections, and it is therefore impossible to determine whether the alignment of the joints corresponds to q-, s-, or d-directions. That the 9°, 140° and possibly 170° sets are of early formation and that they were due to tensional forces is shown by their universal exploitation by the aplitic and hydrothermal fluids, and this, according to Balk (1937), indicates the q-direction rather than the s-direction. However, the alignment of these older joints roughly along the N.W.-S.E. direction indicates their concordance with those so prominently developed in the pre-Cape Granite plutons. In these the dominant N.W.-S.E. direction parallels both the linear

flow structure and the direction of elongation of the plutons and, although they are occasionaly mineralised, apparently represent s-directions (Scholtz, 1946).

The older aplite-bearing veins may follow true q- or cross joints, while the hydrothermal minerals appear to have been emplaced along slightly younger s- or longitudinal joints, these structures probably originating under the regional stresses that were operative on the granites during their intrusion and consolidation. The younger sets, displaying a tendency for preferred orientation in the N.E.-S.W. direction, also have counterparts in the granites, and being barren, succeeded the introduction of the hydrothermal fluids.

IV. THE PETROLOGY OF THE BASIC ROCKS

The essential minerals present are: plagioclase, hypersthene, diopsidic augite, hornblende, biotite, orthoclase and quartz, while magnetite, ilmenite and apatite are the most abundant accessory minerals. Unaltered olivine is not found in any thin section but its presence in the darker varieties of gabbro is suspected and confirmed by chemical analysis. Chlorite, epidote, calcite, saussurite, uralite, talc and serpentine are present as alteration products. The modal proportions of these minerals are tabulated in Table II.

The texture of the rocks is very variable. In the gabbros it ranges from subophitic to monzonitic and poikilitic, and the preferred orientation of the plagioclase and pyroxene crystals in the banded varieties has already been described. The less basic rock-types, the diorites, develop a more uneven, eugranitic texture.

Except in the immediate vicinity of the crush-breccias there is no evidence that these rocks have been regionally deformed by either crushing or shearing. While the extensive alteration of the gabbros may be attributed mainly to weathering processes, to which they display a marked susceptibility, a certain amount of late, hydrothermal alteration has affected the diorites. Thus, although the freshest possible samples were collected during quarrying operations at the harbour, these rocks all showed varying degrees of alteration.

Plagioclase is one of the most abundant minerals in all the rock types and two generations of this mineral are present. The earlier generation varies in composition from an acid labradorite to a basic andesine, while the younger generation, confined to the transitional and dioritic suites, is usually oligoclase.

In the gabbros the individual crystals are distinctly lath-shaped with an average length of 2·5 mm. and a width of 0·5 mm. in the most basic varieties, but slightly less in the poikilitic types. Often the laths show euhedral borders and occasionally develop subophitic intergrowths with the pyroxene crystals, but usually a certain amount of mutual interference of growth between individual plagioclase crystals has resulted in more irregular boundaries. In the presence of large amounts of interstitial orthoclase, slight resorbtion has occurred, the sharp crystal corners having been rounded off and the faces scalloped by the orthoclase groundmass. This feature is most noticeable in the poikilitic varieties, but even here the lath shapes are retained and resorbtion is apparently less than that found in the other rock suites.

The plagioclase feldspars of the gabbroic suite are wholly of the first generation. Their anorthite content ranges from 57% in the most basic banded varieties, to 43% in the poikilitic type, while the average variation in any particular section was never found to exceed 4% An. The anorthite content is further found not to fluctuate

appreciably across any particular band. Zonal structures are generally absent, although a slight normal zoning of less than 2% An. is observed in some of the larger, more basic crystals.

Far greater diversity is displayed by the plagioclase of the rocks of the transitional and gabbroic suites. Seldom do they reach the same dimensions as those of the gabbros, their length ranging from less than 0·1 mm. to about 1·0 mm., while the occasional crystal may exceed 1·5 mm. The largest crystals, with a core composition of from 40% to 49% An., are always the more basic and may present euhedral boundaries to pyroxene or other plagioclase crystals, but their edges against orthoclase or micropegmatite are always highly resorbed and completely anhedral. Except along their euhedral boundaries these crystals always possess a thin, zoned outer layer of more sodic composition. This mantle is so thin that the determined average composition of 28% An. is uncertain.

The smaller plagioclase crystals are universally associated with micropegmatite in which they occur as small rounded grains averaging less than 0·2 mm. in diameter. These show a normal zoning from core to mantle and their composition corresponds to the outer zone of the larger crystals. Their core composition may fluctuate between 38% and 28% An., while their mantles contain between 20% and 28% An.

The two generations of plagioclase feldspar in the rocks of the transitional and dioritic suites are thus clearly apparent. The first generation, represented by the more calcic cores of the larger crystals, corresponds in composition to the feldspars of the gabbros, while the thin outer layers of these and the smaller granular crystals, characterised by a more sodic composition, belong to the second generation. The role of the micropegmatitic fluid as both a resorber of the plagioclase crystals of the first generation and the medium from which the second generation crystallised is indicated.

On the whole the plagioclase is remarkably free from alteration. Only in the vicinity of the aplite and hydrothermal veins has it been much altered, while in some of the "inclusion-bearing diorites" the first generation cores have been rendered almost opaque by minute flakes of secondary material although the second generation mantles and granules remain perfectly transparent.

The relative frequency of the twinning laws exhibited by the plagioclases of the first generation (composition range 43% to 57% An.) in 50 crystals selected at random and determined on the Universal Stage is as follows:—

Albite	52%
Carlsbad	18%
Roc Tourné ...	18%
Albite-Ala	8%
Acline	4%

The twin lamellae of the second generation were found too narrow to determine the feldspar by the usual Universal Stage methods. Albite twins were therefore selected as far as possible and the compositions were deduced from the extinction angles against (010) in the [100]-zone.

Olivine was not found in an unaltered form, but in one section of a particularly dark gabbro abundant rounded grains of highly altered material are believed to be the decomposition products of olivine in view of their frequent inclusion within large host crystals of pyroxene. The average size of these decomposed patches is 1 mm., but those

which are not surrounded by pyroxene may have an average major diameter of 3·5 mm. Three separate alteration products are recognised in these particles:—

(*a*) The central patch is composed of fine colourless flakes of talc which has replaced the olivine in such a way that all the flakes possess an approximately common optical orientation.

(*b*) Fine, dusty particles of opaque iron ore mark the approximate boundaries of the original crystals which sometimes show the typical six-sided form of a prismatic section of olivine. This secondary iron ore is also concentrated along lines which appear to be the traces of the irregular fracture surfaces in the original olivine crystals.

(*c*) Radiating out from the particles are pale green antigorite flakes. This product never occurs within the boundary marked by the outer iron ore layers, but completely surrounds the particles with layers from 0·1 to 0·3 mm. thick and from these it pervades along cracks and cleavages in the other minerals of the rock. It seems probable then that these decomposed patches represent altered crystals of an olivine with a fairly high tenor of iron. The pyroxenes were not found to show similar alteration.

Orthopyroxene is limited to the rocks of the gabbroic and transitional suites where it occurs as small subhedral grains often surrounded by clinopyroxene. The crystals range in size from 0·5 to 2·5 mm. and frequently show alteration to a pale green to colourless chlorite, decomposition commencing along the margins and cleavage traces of the mineral. When fresh, their negative optic axial angle varies between 49° and 52°. According to Winchell (1951) the composition of the orthopyroxene probably ranges from $En_{45}Fs_{55}$ to $En_{39}Fs_{61}$. The pleochroism: α = pale pinkish brown, β = colourless to pale yellow brown, γ = colourless to pale green; and absorbtion: $\alpha > \beta = \gamma$.

Clinopyroxene is the most widely distributed mafic material occurring in all the rock suites. The individual crystals are largest in the gabbros where basal sections often measure up to 4 mm. and prismatic sections up to 6 mm. Their crystal boundaries against the alkali feldspars are complete and euhedral, but a slight rounding of the sharp crystal corners may be detected in the poikilitic monzo-gabbro. Most of the clinopyroxene crystals are riddled with small poikilitic inclusions which may form up to 7% of the volume of the crystal and include orthopyroxene grains, rounded blebs of plagioclase, apatite and iron ore. The many biotite flakes occurring in the pyroxenes are regarded as "advance islands" in view of their exploitation of the cleavage directions and the fact that all the grains in a single pyroxene crystal show a common optical orientation. Similarly, adjacent interstitial orthoclase frequently veins the pyroxenes along the open basal parting or as irregular stringers aligned parallel to the prominent prismatic cleavages, and isolated blebs of this feldspar, showing optical continuity with the surrounding orthoclase ground, are taken to be transverse sections of such veins rather than poikilitic inclusions.

The same mineral is also abundant in the dioritic and transitional suites but, in the former, comparable dimensions are never obtained and the grain size is almost always less than 1 mm. In the dioritic suite too, the crystals are often resorbed and surrounded by reaction rims of hornblende and in some hornblende crystals only small skeletal remnants of pyroxene remain. This reaction is not common in rocks of the transitional zone.

The mineral appears to be a colourless to pale green diopsidic augite or endiopside, characterised by the following optical constants:—

$2V\alpha = 51°\text{-}52°$; $\gamma/c = 38°\text{-}39°$

$\beta = 1{\cdot}673 \pm 0{\cdot}002$; $\gamma - \alpha = 0{\cdot}0235$

Dispersion weak, $\rho > v$

According to Deer and Wager (1938) this represents an augite with composition $Wo_{38}En_{55}Fs_7$. It is interesting to note that a pyroxene with similar optical constants is found in the quartz-hyperites of Insizwa (Scholtz, 1936) and has been described in the rocks of the Basal Zone of the Bushveld Complex. These pyroxenes, however, usually display diallage parting parallel to (100), but this feature is not observed in the augites from Yzerfontein, although twinning, both simple and polysynthetic, always has this direction as composition face.

Zoning of the clinopyroxenes is rare. In two crystals, however, a slight zoning was observed whereby the cores have slightly lower axial and extinction angles ($2V\alpha = 49°$; $\gamma/c = 37°$) than the mantles which possess optical constants corresponding to those given above.

Like the orthopyroxene, the augite may alter to a pale green chlorite, but alteration to flaky uralite is also common.

Primary Hornblende is limited to the rocks of the dioritic suite, of which it may constitute up to 17%. The mineral almost always occurs as reaction rims surrounding pyroxene granules and the boundary between these two minerals is irregular and intergrown. The hornblende is usually confined to an outer reaction envelope but may penetrate far into the pyroxene crystal along cleavage and parting directions, and isolated patches of hornblende in common orientation with the material forming the reaction rim may riddle the larger pyroxene crystals. The reaction rims are further seldom homogeneous but contain many poikilitic inclusions of similar shape, size and mineral constitution as those found in the pyroxenes. Rarely are hornblende crystals found which do not contain pyroxene cores, and they may themselves be surrounded by reaction rims of biotite.

All the hornblende is of a green variety the optical constants of which are as follows:—

Optic axial angle:	$2V\alpha$	$= 76°\text{-}82°$
Extinction angle:	γ/c	$= 12°\text{-}16°$
Pleochroism:	α	= pale straw
	β	= pale green to olive-green
	γ	= brownish green, rarely bluish green
Absorption:	$\alpha < \beta < \gamma$	

Biotite is a component of all the Yzerfontein rock types constituting from 1·5% to 10% by volume.

The mineral is most commonly a golden brown type but may show slight variation of optic properties. It is probably a siderophyllite as shown by the following optical constants:—

Pleochroism:	α	= pale brownish yellow
	$\left.\begin{matrix}\beta\\\gamma\end{matrix}\right\}$	= golden reddish brown to blackish brown
Absorption:	$\alpha < \beta = \gamma$	
Refractive indices:	$\gamma = 1{\cdot}641 \pm 0{\cdot}002$	
	$\alpha = 1{\cdot}600 \pm 0{\cdot}002$	

The biotite occurs abundantly as a reaction product of pyroxene and hornblende. In the quartz-free rocks it forms directly by reaction with pyroxene, but in the presence of micropegmatite or quartz the pyroxene is first replaced by hornblende which is in turn converted to biotite. The mineral is also often associated with the accessory minerals, apatite and ilmenite, or it may occur as a primary mineral in the form of large equidimensional flakes. Further, in the poikilitic gabbros, nodules composed of small, radiating and bent flakes of biotite are common in the large orthoclase individuals.

The mineral commonly alters to a pale green to colourless chlorite, usually with the segregation of fine ore particles along the cleavages.

Alkali Feldspar and Quartz occur both as micropegmatitic intergrowths and as individual minerals.

In the gabbros, alkali feldspar is typically interstitial and forms the groundmass in which the other earlier crystals, with idomorphic outlines, are embedded. Its appearance is that of a late liquid filling the spaces between the components of a relatively porous crystal mush, and for large stretches this feldspar ground mass shows optical continuity. The mineral constitutes varying proportions of the gabbro and Table II shows that it is most abundant in the poikilitic varieties. The figure here given is, however, considered misleadingly high as the cut sections usually covered only the large poikilitic orthoclase insets and seldom included the orthoclase-free gabbro between these portions.

The alkali feldspar of the gabbro is always perfectly fresh and either untwinned or simply twinned according to the Carlsbad law. The optical axial angle, $2V\alpha$, varies from 61° to 66° indicating soda-orthoclase. The optical constants of the alkali feldspar in the micropegmatite are more variable but also indicate a soda-orthoclase.

Micropegmatitic intergrowths of alkali feldspar with quartz are limited to the rocks of the dioritic and transitional suites. Here too the mineral has the appearance of having been introduced as a late fluid which bore a reaction relation to the minerals of earlier crystallisation. The components of this intergrowth may show optical continuity over large areas or they may form smaller shapeless grains with diameters ranging down to about 0·2 mm. The intergrowth has by no means been evenly developed, most of the grains showing portions entirely free of quartz. Where it does exhibit homogeneous development the ratio of quartz to feldspar was measured and found to remain practically constant at 27·2% : 72·8%. The quartz stringers of the micropegmatite are seldom wider than 0·02 mm. and under normal magnification they give the micropegmatite a dusty appearance. This is also due to the presence of many fine inclusions, most notably long slender needles of apatite with average dimensions of 0·04 by 0·005 mm. Among the larger inclusions apatite and iron ore are prominent.

Just as orthoclase free from quartz may be found in the transitional and dioritic suites, so also are patches of free quartz found in these rocks. The mineral is, however, usually subordinate and fills interstitial voids between the minerals of earlier crystallisation.

The Accessory Minerals are apatite, magnetite, ilmenite and sphene. The apatite of the gabbro occurs as fairly large euhedral crystals and is present as inclusions in practically all the minerals of this suite. Apatite crystals of a similar size and appearance are also found in the transitional and dioritic suites. In these latter suites there are also developed smaller acicular apatites chiefly associated with the micropegmatite.

Much magnetite has been released by the alteration of the mafic minerals but the primary mineral is more abundant. Together with ilmenite it is prominently included in the biotite and pyroxene grains. Small amounts of sphene are found in the less basic rocks, while zircons are absent. Pyrite, tourmaline, epidote, calcite and hematite may be developed in accessory amounts by hydrothermal alteration.

V. REACTION RELATIONS IN THE DIORITES

Reaction rims are commonly observed amongst the dark minerals of all the Yzerfontein igneous rocks, their best development being attained in the dioritic rock (augite-biotite-diorite) collected at the Vlaekop beacon where resorbed cores of augite are completely surrounded by hornblende, which may in turn have rims of biotite. A further example of a discontinuous reaction relation is shown between the hyperthene and the surrounding augite in the gabbroic rocks, and an example of a continuous reaction series, by the normal zoning of the plagioclase crystals in the diorites.

During the petrological examination of the rocks it was observed that, when a twinned crystal of augite was surrounded by a reaction rim of hornblende, the replacing hornblende was also twinned and in such a manner that the twinning laws were identical and that one and the same plane formed the composition face of the twin in both minerals. This could only be explained by a preferential orientation of the reacting material towards the orientation of the core mineral being replaced. It also indicates, incidentally, that the (100) twinning of hornblende is paragenetic rather than metagenetic. After the measurement of the relative orientations of a number of pyroxene cores and amphibole mantles, constant relation between the two parts became apparent.

An attempt was therefore made to determine whether such constant orientation relations are common between the rim and core material in the discontinuous reaction series of N. L. Bowen (1928, p. 61):—

olivine—Mg pyroxene—Ca-Mg pyroxene—amphibole—biotite.

The determinations were made primarily on the Yzerfontein rocks, but they were supplemented by and checked against observations made on rock-sections from the Malmesbury diorite, the picrites and hyperites of Insizwa, the Bushveld gabbros and norites and rocks from the basal zone of the Stillwater Complex, Montana. The investigation showed that such relations undoubtedly exist, and in Fig. 2 stereographic projections of the principal morphological and optical directions of these minerals have been arranged in such a way as to reflect their mutual orientations in reaction rims.

Olivine-pyroxene reaction.

No observations of this nature could be made on the Yzerfontein rocks, but determinations on other olivine-bearing rocks show that a complex relation may obtain between the two minerals. This becomes apparent when the crystallographic b-directions of a number of pyroxenes are plotted stereographically relative to the b-axes of an olivine crystal. The tendency for the pyroxene b-axes to concentrate around the poles of $\{101\}$, $\{160\}$, $\{1\bar{3}1\}$, etc., is then observed. These directions appear to be related to the face-normals of the isolated Si-O tetrahedra forming the fundamental structure of the olivine.

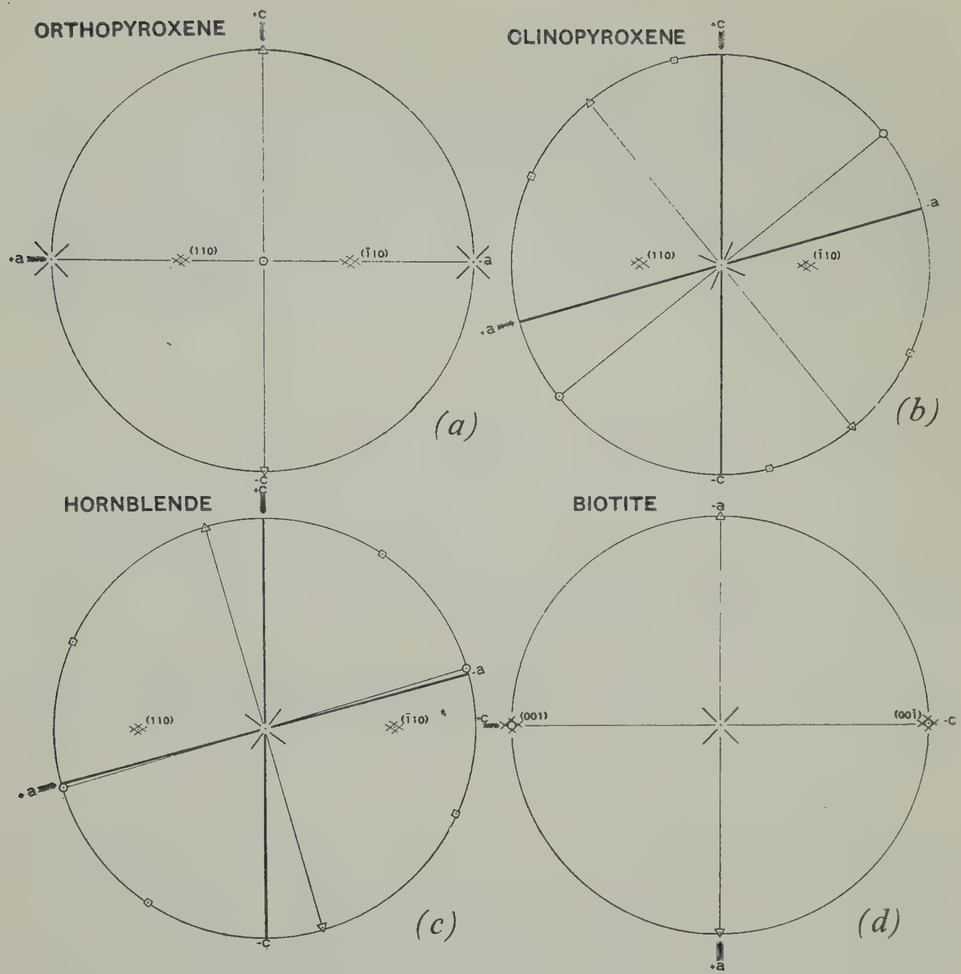


FIG. 2.

Illustrating the relation obtaining between the aether axes (α = ⊙, β = +, γ = △) and the a-and c-crystallographic axes in the (010) plane of the principal dark minerals of the reaction series.

Orthopyroxene-clinopyroxene reaction. (Figs. 2 (*a*) and 2 (*b*)).

Hypersthene is often surrounded by or in contact with augite in the gabbroic rocks of Yzerfontein, but true reaction or replacement structures are comparatively rare. Their inter-relationship suggests rather that the two minerals separated from the liquid more or less simultaneously and the greater part of their crystallisation periods overlapped. Under such circumstances true reaction rims of clinopyroxene about orthopyroxene are not expected to be abundant. Where they are developed, however, the two minerals of the reaction pair show common cleavage traces, while the α-axis of the ortho-pyroxene coincides with the β-axis of the clinopyroxene. Adopting Groth's orientation of the orthopyroxene, such that the axial ratio a : b is

greater than unity, these observations show that the b- and c-crystallographic axes of the replacing clinopyroxene are arranged parallel to the corresponding axes of the orthopyroxene.

A similar relationship was observed by H. J. Nel (1940, p. 48) in the diabases of the Basal Zone of the Bushveld Complex where "monoclinic pyroxene is sometimes found as a mantle round the hypersthene, the latter in this instance always exhibiting embayed outlines". J. C. Boshoff (1942) reports a like feature in the diallage-gabbros of the Upper Zone of the same intrusion, and he shows further that, where the orthopyroxenes of the olivine-diorites exhibit sets of fine, clinopyroxene exsolution lamellae, the same orientation is common but not exclusive. G. S. J. Kuschke (1939, p. 62) and H. J. Nel (1940, p. 54) confirm this for rocks of the Basal and Critical Zones of the Complex.

Clinopyroxene-Amphibole reaction. (Figs. 2 (*b*) and 2 (*c*)).

In every case of the development of hornblende reaction rims about clinopyroxenes at Yzerfontein, the following mutual relations are found:—

(*a*) the optic axial planes of the two minerals are parallel,
(*b*) the prismatic cleavages of the two minerals are tautozonal.

Since the β-axis of each mineral is coincident with the b-crystallographic axis of that mineral, and since the cleavages in each case lie in the (100) zone, it follows that the amphibole replaces the pyroxene in such a way that their b- and c-crystallographic directions coincide. The a-axes, however, differ slightly, this being due to the slight difference of monoclinic angle of the two minerals.

In the investigation it was also observed that the direction of one of the optic axes of the augite always coincides with one of the hornblende, while the direction of the other augite optic axis very nearly parallels the γ-direction of the hornblende. This relation is, however, probably purely coincidental, resulting from the particular optic constants of these two mineral varieties, and is not expected to be general.

An attempt to trace literature on this point revealed that, as far back as 1896, A. Harker (1896, p. 324) described similar relations where augitic xenocrysts derived from gabbro are surrounded by reaction rims of green hornblende in the granophyres of Strath (Skye), replacement occurring in such a way that the b- and c-axes are common to both portions. It is perhaps interesting to note that these two minerals have identical extinction angles to those found at Yzerfontein.

In his description of the diorites of Klein Paardeberg, Malmesbury district (Oranjefontein stock), A. W. Rogers (1905, p. 48) mentions that "the pyroxene sometimes forms complicated intergrowths with the hornblende and also occurs in the centre of large hornblende crystals; in such cases one set of prism cleavages is common to both minerals". Microscopic examination of the rocks from this area, however, failed to disclose such relations, but fully confirmed the replacement noted at Yzerfontein.

Amphibole-biotite reaction. (Figs. 2 (*c*) and 2 (*d*)).

The examination of the orientation of biotite reaction rims relative to that of hornblende cores shows that:—

(*a*) the optic planes of the two minerals are parallel,
(*b*) the biotite cleavage (001) is parallel to the (100) direction of hornblende,
(*c*) α of biotite falls in the [001]-zone of hornblende,
(*d*) the γ-axis of biotite parallels the c-axis of hornblende.

Resolving these to the morphological directions of the two minerals, biotite replaces hornblende in such a way that their b-crystallographic axes remain common, while the a-direction of biotite coincides with the c-direction of hornblende. In addition, this arrangement preserves the tautozonal disposition of the prominent cleavages of the two minerals, a feature already noted in the pyroxene-amphibole reaction.

Occasionally, biotite may appear as a direct reaction product with pyroxene and here the relative alignment of the two minerals is the same as it would have been had an intermediate reaction layer of hornblende been present. D. L. Scholtz (1936, p. 116) noted this tendency towards parallel alignment of the aether axes of biotite flakes replacing pyroxene in the hyperites of Insizwa.

To summarise, it appears then that when an earlier mineral of the discontinuous reaction series:—

olivine-orthopyroxene-clinopyroxene-amphibole-biotite

is surrounded by a later one of the same series, *a definite crystallographic relation* obtains between the two individuals. When olivine forms the core, the relations may be complex, but in all the other cases the b-crystallographic axes coincide, while the most prominent cleavages tend to be tautozonal.

The above associations may relate to the conversion of the fundamental silicon-oxygen tetrahedral structure of the mesosilicates, first to that of the inosilicates and then to the phyllosilicates with decreasing temperature. The relations are such as to require the least possible rearrangement of this fundamental structure, while the diffusion of the ions not forming part of the silicon-oxygen framework would accompany such conversion. Such features demand equal freedom of growth in all directions and would therefore be expected to obtain only under the influence of hydrostatic pressure, such as would exist in a fluid, magmatic environment, but not in the restricted, solid state.

VI. THE ANALYSIS OF THE PRIMARY BANDING OF THE GABBROS

The poverty of exposures of rocks of the gabbroic suite renders impossible the task of determining the probable form and size of the gabbro body on the distribution of these outcrops alone. The development in these rocks of primary rhythmic banding was, however, believed to have some genetic significance, and in an attempt to obtain additional information this feature was investigated in greater detail.

The field appearance of the banding has already been described and in this connection it simulates both the "rhythmic layering" of the Layered Series of the Skaergaard Intrusion as described by L. R. Wager and W. A. Deer (1939), and the "primary banding" observed by H. H. Hess (1938) in the norites of the Stillwater Complex.

Mineralogically the banded rock is essentially composed of varying proportions of labradorite, augite, hypersthene, biotite and orthoclase. The plagioclase and pyroxene usually occur as large, individual, well developed crystals, although rounded grains of orthopyroxene may often be included in more perfect clinopyroxene crystals. Their appearance may suggest their accumulation as discrete crystals and, solely for descriptive purposes, they may be called the early precipitate, while the interstitial orthoclase and biotite appear to have crystallised subsequently from a liquid which surrounded the early crystals and may be termed the rest magma or inter-precipitate material. Thin sections cut across the plane of banding exhibit a directed texture as

the crystals of the primary precipitate appear to be preferably orientated with their largest dimensions parallel.

This feature is described as igneous lamination and in order to ascertain its genetic connection with the primary banding, these rocks were subjected to a series of petrofabric analyses. The direction followed by this investigation was an examination of the frequency of the orientation of the primary crystals in thin sections cut both parallel to and at right angles to the banding. Following custom, the three mutually perpendicular fabric axes have been chosen so that the c″-axis is normal to the banding and the a″- and b″-axes lie within this plane. The b″-axis is horizontal and corresponds to the strike of the banding at 72° M.N. while the a″-axis dips towards the south-east at an angle of 68°. The density gradation indicates that the base of the body lies towards the north.

The results of the investigation of clinopyroxene and plagioclase are, for convenience, discussed separately.

(1) *Clinopyroxene.*

Most of the crystals of this mineral show a distinct elongation in the direction of the c-crystallographic axis and, in addition, show flattening parallel to the clinopinacoid crystal face. Contoured petrofabric diagrams showing the concentration of these features in various orientated sections were therefore prepared after the method described by Fairbairn (1942).

Diagrams showing the distribution of the c-crystallographic axes all display a pronounced concentration of points along a belt seldom more than 20° broad and coincident with the a″b″-fabric plane. This is clearly illustrated by Fig. 3 (*a*), a collective diagram of information obtained from three sections cut perpendicularly to the c″-fabric axis. The complete girdle concentration along the periphery of the projection circle indicates that the c-crystallographic axes of the clinopyroxene crystals either lie in the plane of banding or are inclined thereto at a slight angle. The slight grouping of poles at two points on opposite sides of the circle, constituting a maximum in this plane, may indicate minor directional control over the alignment of the grains.

Closely related to this pattern is the pronounced axial symmetry of diagrams showing the distribution of β-aether axes of this mineral. Fig. 3 (*b*), prepared in a similar way to Fig. 3 (*a*), clearly illustrates this concentration of β-axes approximately parallel to the c″-fabric direction. Since the direction of β coincides with [010] or the normal to the (010) crystal-face, the arrangement of the clinopinacoidal faces of clinopyroxene parallel to the plane of banding is indicated by this pattern.

Single crystals of different orientation from those described above were occasionally encountered but, through limiting the lowest contour value to 2%, they do not appear in Figs. 3 (*a*) and (*b*).

(2) *Plagioclase Feldspars.*

The plagioclase crystals of the banded gabbro generally possess a tabular habit, "a habit which is usual for bytownite and labradorite in basic igneous rocks" (Wager and Deer, 1939). Cross-sections of these crystals, however, are nearly always rectangular and flattening is commonly parallel to the clinopinacoid, but larger faces may occasionally be developed in the case of (001), (021), ($0\bar{2}1$), etc. Because of this, it was considered inadvisable to express the orientation of the plagioclase crystals in terms of their crystallographic directions. Instead, only the orientations of the longest axes of all feldspar laths showing a fairly pronounced elongation were measured in sections cut in the three principal fabric directions. The results so obtained were

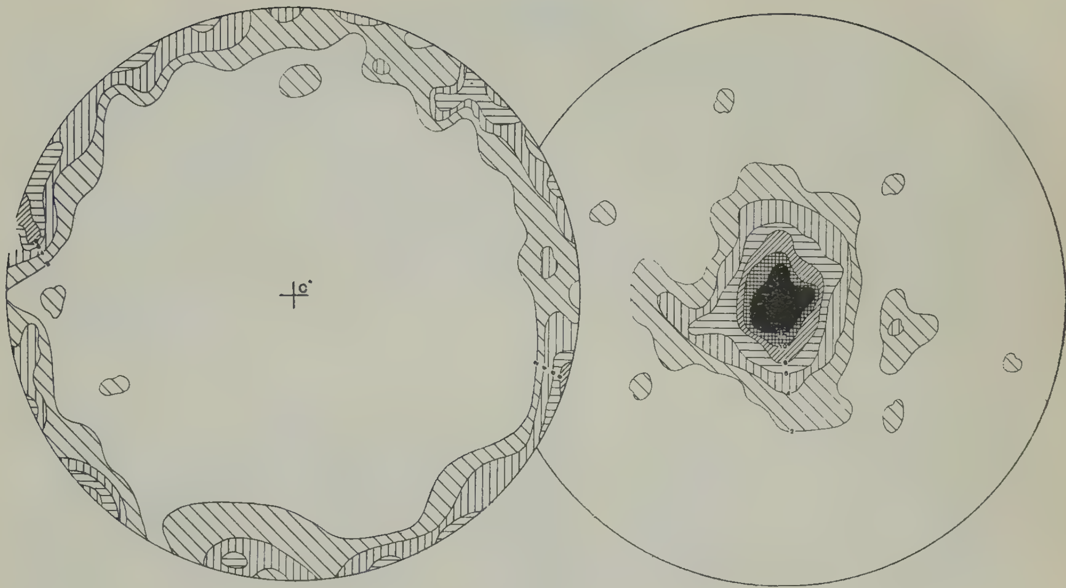


FIG. 3.

(*a*) (*b*)

Petrofabric diagrams of sections cut perpendicular to the c″-fabric axis, showing the distribution of (*a*) 100 c-crystallographic axes, and (*b*) 100 β-axes of clinopyroxenes. Distribution contours at 2-4-6-8-10-12%.

utilised in preparing azimuth diagrams as described by Fairbairn (1942). On the one hand, diagrams of sections cut perpendicular to the c″-fabric axis show a random orientation of the elongation of these crystals in the plane of banding, while on the other hand, Fig. 4, which is of sections cut parallel to the c″-axis, indicates a pronounced common orientation of the elongated feldspares in such a way that their longest axes lie parallel to the plane of banding.

To summarise, the principal features of the igneous lamination of the banded gabbros at Yzerfontein are such that the somewhat flattened early crystals all have their greatest dimensions parallel to the plane of banding. In petrofabric diagrams this is apparent as a complete belteroporic girdle concentration of the elongation elements perpendicular to the c″-fabric axis, and a pronounced axial symmetry of the normals of the best-developed crystal faces parallel to this direction.

According to E. B. Knopf (1938, pp. 70 and 123), axial symmetry is characteristic of the depositional fabric of non-tectonites, but may also be developed in *B*-tectonites. Similarly, a girdle fabric may be displayed by both tectonites and non-tectonites, but, whereas the non-tectonite depositional girdle is normal to the c″-axis, tectonite girdles usually parallel this direction. The preferred orientation of the crystals of the Yzerfontein banded gabbros therefore indicates a depositional fabric, which results from "the deposition of particles in a stagnant medium by a settling movement under the influence of gravity." (Knopf and Ingerson, 1938.)

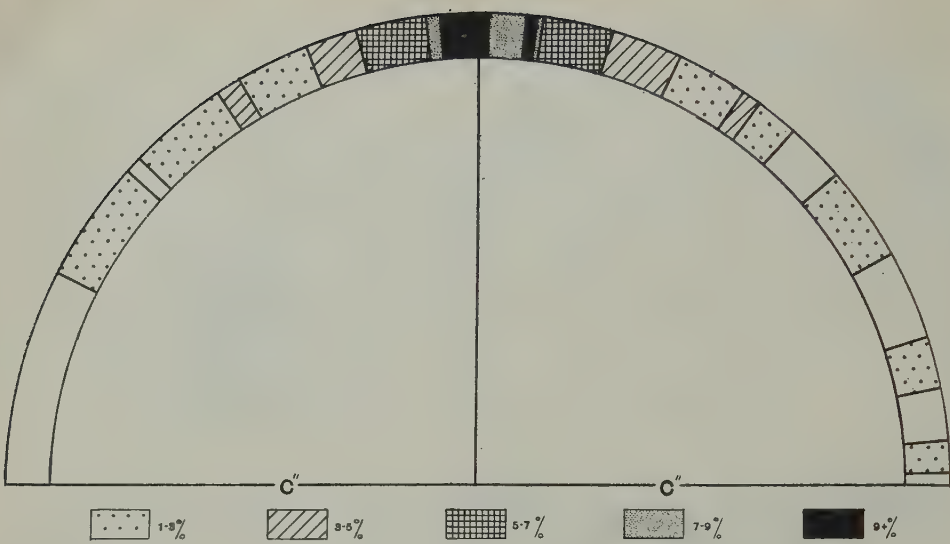


FIG. 4.

Azimuth diagram of the orientation of the elongation of 200 feldspar laths, in sections cut parallel to the c″fabric axis.

Various theories to explain the formation of primary rhythmic banding have been advanced and have been ably reviewed by R. A. Daly (1933), R. R. Coats (1936), H. H. Hess (1938), and L. R. Wager and W. A. Deer (1939). These theories may now be briefly referred to especially on the basis of the fabric of igneous lamination, in order to ascertain which would best explain the rhythmic banding as observed at Yzerfontein. Obviously, in view of the clear depositional fabric, any process operative in the solid state could not have obtained.

The magmatic theories may be broadly divided into the following:—

(*a*) theories which require the intrusion of two different magmas,
(*b*) theories relying on strong convection currents in the magma during crystallisation, and
(*c*) theories of gravitative settling from a quiet magma.

(*a*)

Although the proportions of the minerals vary across the individual bands, they show no variation in either composition or size. This is suggestive of a consanguineous relationship between even the most extreme rock types in the banded gabbros, and the banding could therefore hardly have resulted from the intrusion into the same chamber of two different magmas, either simultaneously as proposed by A. Harker (1904), or successively as suggested by N. L. Bowen (1928). The regular rhythmic nature of the layering and the well-defined density gradation across the bands would also be difficult to explain by such a process.

(*b*)

F. F. Grout (1918) has suggested that convection currents in the magma were responsible for the banding of the Duluth gabbros, while Wager and Deer (1939)

conclude that a similar process, coupled with crystal fractionation, resulted in the primary layering of the Skaergaard Intrusion. In the latter instance, crystals of early precipitation which formed at the roof and margins were transported to the floor of the intrusive body where they were graded according to their densities and evenly distributed by these strong magmatic currents. The rhythmic nature of the banding is attributed to the periodic nature of these currents. Much evidence from the field shows the existence of the powerful convection currents, the most important being the distinct linear arrangement of the rectangular plagioclase laths in the plane of igneous lamination. The absence of any pronounced lineation at Yzerfontein emphasises rather the quiescence of the gabbro magma during the formation of the bands, and crystal settling, without the aid of convection currents, seems a more likely explanation of this feature.

(*c*)

The earlier theories of crystal settling of P. A. Wagner (1924), A. L. Hall (1932) and R. A. Daly (1933) were more in the nature of preliminary suggestions and were conceived to explain the large-scale, regional banding observed in the Bushveld Complex rather than the type of local, rhythmic banding under consideration. R. R. Coats (1936) first advanced the theory that rhythmic banding could result from a process involving the simultaneous crystallisation and settling under the influence of gravity of two or more minerals in a magmatic medium of lower density. This mechanism of "rhythmic differential settling" was, moreover, supported by laboratory experiments.

An essentially similar theory has been advanced by H. H. Hess (1938) to account for the banded norites of the Stillwater Complex. The rhythmic nature of the banding, as well as the occurrence of "zones of disturbed banding", was explained by "disturbance of the normal state of quiescence in the magma by short epochs of mild, but irregular turbulence". In both hypotheses the actual deposition of the crystals takes place from a practically motionless magma and igneous lamination, with a distinct non-tectonite depositional fabric, may therefore be expected to be developed by either process. On the texture alone then, neither mechanism can be favoured, but the presence of possible zones of disturbed banding at Yzerfontein may indicate that the process outlined by Hess has here been operative.

A recent petrofabric study by J. J. van den Berg (1946) of the igneous laminated gabbros of the Bushveld at Bon Accord has revealed a fabric so similar to that of the Yzerfontein banded gabbros that a comparison of the two is instructive. Orthopyroxene and plagioclase crystals form the early precipitate of the Bon Accord gabbros, and the preferred orientation of these crystals gives the rock a fairly distinct non-tectonite depositional texture. The formation of these rocks by fractional crystallisation and crystal settling, as first proposed by P. A. Wagner (1924), is thus fully substantiated. In an important respect, however, the fabric of these rocks differs slightly from both the ideal depositional fabric, as described by E. B. Knopf (1938), and that of the Yzerfontein gabbros. This is apparent in petrofabric diagrams of the elongated elements which show a fairly definite preferred alignment parallel to the strike direction of the igneous lamination. Van den Berg interprets this feature as the effect of "the force of intrusion" of the magma on the settling crystals. J. Ellis (1946) suggests that the lineation is due to glide and recrystallisation in the solid phase.

Neither of these explanations appears to be satisfactory, the former because the "force of intrusion" during crystallisation was the isotropic hydrostatic pressure of the fluid magma, and the latter in view of the non-tectonic nature of the fabric.

The typical girdle as postulated by E. B. Knopf is, however, such as would be formed only by deposition on a flat floor, and the lineation observed by van den Berg could more suitably be explained by a slight variation of the normal crystal settling process.

The non-tectonic depositional fabric may be attributed to two factors: the planar control of the floor of deposition, and the linear gravitational force of settling; and for perfect development it must be assumed that the latter acted perpendicularly to the former so as not to destroy its planarity. If, however, the floor of deposition slopes, then a component of gravity would act in the direction of the dip of the floor and the linear elements would show some degree of preferred orientation within the plane of foliation and at right angles to this component. This lineation would disturb the perfection of the even girdle fabric pattern by displaying a maximum concentration of elongation at right angles to the slope of the floor, while not appreciably affecting the axial symmetry of the tabular elements.

Unfortunately, literature pertaining to the petrofabrics of undoubted igneous non-tectonites is surprisingly limited and no mention of the variation of the attitude of the floor of deposition could be found.

If, however, the above argument is correct, then the fabric of the Bushveld gabbro at Bon Accord is exactly that which would be expected from the settling of crystals under the influence of gravity on a *moderately inclined* floor. Although the floor of the Bushveld is generally believed to be basined, and around the periphery both the planes of pseudostratification of the basal rock and the contact with the underlying sediments dip regularly towards the centre, often at high angles, practically all writers on the Complex apparently consider that the attitude of the floor and the pseudostratification was horizontal or nearly so, not only at the time of the intrusion of the basic magma, but also during its consolidation. The crustal collapse which caused the central subsidence is generally regarded as having been initiated by the intrusion of the later Red Granite magma into a position above the solid norite body. No positive evidence has, however, been submitted which would justify this insistence on an initial horizontal, rather than a slightly basined attitude of the norite body.

Thus, P. A. Wagner (1924, p. 77 and 1929, p. 41), on the presumptive evidence that the rocks of the Transvaal System, forming the floor of the body, were horizontal or nearly so at the time of the basic intrusion, accepts unreservedly G. A. F. Molengraaf's original suggestion (1904, p. 57) and assumes that the present basin-like disposition of the Complex originated subsequent to the consolidation of the igneous rocks composing it. A. L. Hall (1929, p. 12 and 1932, p. 201 *et seq.*) relies on the horizontal attitude of the norite body during its consolidation to account for the remarkable conformity of the plane of pseudostratification displayed by the basic rocks, and concludes that the original norite body had the form of "a gigantic sub-horizontal sill". A. L. du Toit (1939, p. 174), also impressed by the regularity of the mineral banding, contends that the pseudostratification "was produced in an approximately horizontal position" and that the basining was presumably subsequent to the consolidation of the norite.

A notably different view is, however, expressed by R. A. Daly (1928), who indicates the significance of simultaneous crustal subsidence and magmatic intrusion, as by this means the depression of salic material to a hotter region may have been responsible for the generation of the younger granite magma. Daly therefore questions the majority opinion that the banding was initially horizontal and, by analysing the mechanics of the intrusion, believes that "the magmatic extrusions meant the transfer of material from beneath the crust, which was therefore, inadequately supported", and demonstrates that the central subsidence and basining of the Transvaal beds was

closely connected with the emplacement of the Complex. While the weight of igneous material on the sedimentary floor must necessarily have functioned in an analogous manner, Daly apparently regards its effect as being far less important.

The close relationship between the attitude of the Bushveld Complex and the rocks of the Transvaal System cannot be overemphasised and there can be little doubt that the intrusion of the Bushveld magmas was an integral phase of the geosynclinal cycle of the deposition of the Transvaal sediments. Unfortunately, the sedimentational aspects of these rocks have not yet been fully investigated, but the primary structures indicate that deposition frequently occurred under shallow water. In order to accommodate the considerable thickness of sediments (varying between 9,000 and 16,000 feet) and yet retain a shallowly submerged upper surface of deposition, continual gradual subsidence of the sediment-collecting basin must have kept pace with the deposition.

On the one hand then, there is no reason to believe that this sagging ceased during or immediately following the intrusion of the magma, especially as the emplacement of the Complex accompanied the foundering of the crust, while on the other hand, justification could not be found for the argument that the regular banding could only have been formed by differential crystal settling on a horizontal floor, and the development of equally distinct banding with an original, moderate inclination is believed to be possible. Under the circumstances, it may reasonably be deduced that the floor of deposition of the basal rocks of the Bushveld Complex possessed an initial shallow, though somewhat elongated, basin-like form and that the present high centripetal dips recorded along the margin of the Complex must not be attributed entirely to conditions closely following the intrusion of this magma, as considerable central subsidence followed the consolidation of these rocks. This must naturally be attributed to the adjustment of isostatic equilibrium, necessitated by the tremendous load carried by this portion of the crust. Du Toit (1933, p. 11) shows that the latest movement in this direction dates to as recent a time as the mid-Tertiary and amounted to a sinking of the surface of the basin of perhaps 1,000 feet. The evidence for this warping has been deduced largely from the behaviour of the diamondiferous gravels of the Lichtenburg and Ventersdorp district when traced up to the margins of the depression (du Toit, 1933), and the nature of the antecedent rivers which cross the Bushveld Basin of the Central Transvaal (Maske, 1956).

From the above discussion we may conclude that the sedimentary floor of the Bon Accord gabbros, at the time of their intrusion, was moderately inclined, although the present centripetal dip of 30° must be attributed in part to considerable post-consolidation tilt. The fabric of the rock indeed seems to substantiate this conclusion and it is reasonable to assume that the development of the lineation observed in such a fabric is a function of both the degree of inclination of the floor and the elongation index of the crystals. In this respect it is interesting to note that in a preliminary investigation (based on measurements of only two thin sections) of the fabric of pyroxenites from the Eastern Transvaal, van den Berg (1946) observes that there is no evidence of linear orientation owing to the equidimensional habit of the pyroxene crystals. The elongation index of the orthopyroxenes of the Bon Accord gabbro has an average value of 3, while that of the clinopyroxenes of the Yzerfontein gabbro is but slightly less. The preferred orientation of these elements in the plane of foliation is, however, decidedly inferior in the latter when compared to that displayed by the former, and it is believed that the inclination of the floor of the Yzerfontein gabbro body could not have exceeded the order of a few degrees during the consolidation of the banded rocks.

The small extent of these rocks, however, renders a deduction of either the shape or the size of the body speculative. The regularity of the banding and the coarseness of grain suggests a fairly large body, probably a thick sill or laccolith, in which, according to Wager and Deer (1939), powerful, regular convection currents are less likely to develop than in a funnel-shaped body. Further, as the process of rhythmic differential crystal settling requires the presence of a "head" of magma above the consolidating rock, the banded gabbro presumably occupied a position in the lower portion of the body. The attitude of the banding is no longer horizontal but dips steeply towards the south-east. This tilting, necessarily subsequent to the consolidation of the layered rock, indicates that the greater portion of the gabbro body should lie beneath the surface on the south-eastern side of the foliation plane of these banded rocks.

In conclusion, it would seem as if the fabric displayed by gravitational accumulations of early minerals of bladed habit enables the petrographer to deduce the initial attitude of the depositional surface of intrusions characterised by fractional crystallisation along a gravity gradient.

VII. THE PHASE PETROLOGY OF THE BASIC ROCKS

Owing to the gradation between the various rock types, the lack of sharp intrusive contacts between them and the poverty of primary flow structures in the less basic varieties, it was considered advisable to map the Yzerfontein Complex according to the distribution of "critical mineral phases" rather than by the traditional practice of mapping arbitrary "rock species". The former method of mapping is attributed to Shand (1942), and the terms "phase" and "critical phase" are here used not in the strict thermodynamic sense but in the sense implied by Shand. Applying this method in a modified form, van Zyl (1950) was able to ascertain the stock-like form of the Malmesbury diorite bodies by mapping the distribution of the dark minerals.

At Yzerfontein, it was found that olivine, hypersthene, hornblende and quartz are critical phases, their distribution being restricted to particular parts of the complex and also to particular periods in the cooling history of the magmas. Although plagioclase is widely distributed in all the rocks, the recognition of two generations of this feldspar, namely, an earlier and widely distributed variety, and a later type characterised by a more restricted occurrence, permits the distinction of a critical oligioclase phase. Quartz is usually microscopically intergrown with orthoclase in the form of micropegmatite, and, since the occurrence of this intergrowth is limited, it is also critical. As the distribution of the micropegmatite is identical to that of quartz, it was found convenient to map the distribution of micropegmatite.

To a certain extent the distribution of these critical phases is interrelated and the mapping of this distribution discloses the existence of three distinct zones (see accompanying geological map, Plate III):—

Z 1. An olivine-hypersthene zone.
Z 2. A hypersthene-micropegmatite-oligioclase zone.
Z 3. A micropegmatite-oligoclase-hornblende zone.

The apparent parallelism between the location of the zones and the distribution of the various rock suites follows, of course, from the method whereby the rocks of the complex were subdivided into their different suites. The ultimate basis for this classification has been mineralogical. Thus Z1 covers the distribution of the gabbroic suite;

Z 2 comprises what has been referred to as the "transitional zone"; while the rocks in Z 3 correspond to the dioritic suite.

Although augite and basic plagioclase feldspar do not represent critical minerals, their ubiquitous distribution is important as it offers a clue to the relative ages of formation of the zones. During their petrological examination, it was noted that in Z 1 these minerals show their most perfect crystalline development and largest dimensions, while resorption, replacement and a decrease of grain size is observed in Z 2 and Z 3. In other words, the formation and growth of these minerals in Z 1 was followed by their corrosion, resorption and replacement in Z 2 and Z 3, thereby implying that the rocks of Z 1 are older than those of Z 2 and Z 3.

The original form and extent of the gabbro intrusion (Z 1) has been discussed and the positions of Z 2 and Z 3 indicate the discordant intrusion of a later magma cutting across the strike of the primary banding of the gabbro. Field and microscopic evidence suggests the incorporation in this magma of much disintegrated gabbroic material, probably derived from both crystalline and partly crystalline xenoliths and xenocrysts. The micropegmatite appears to be genetically related to the later magma which reacted with the augite and basic plagioclase xenocrysts of the gabbro, precipitating the critical hornblende and oligoclase phases, and the resultant, highly contaminated magma appears to have finally crystallised as the diorites of Z 3. Apart from the discordant attitude of the diorite body and its possible downward and eastward extension, phase petrology could offer no further evidence as to the form of this body.

Both the intermediate position of Z 2 and the field characteristics of this zone suggest that it is a marginal phase of the diorite in which further mixing between the gabbro and the contaminated dioritic liquid occurred.

VIII. THE CHEMISTRY OF THE IGNEOUS ROCKS

As no analyses of the igneous rocks of Yzerfontein were available and since chemical data were considered necessary for the interpretation of the relations between the rock types, 14 representative samples were submitted to the Division of Chemical Services for analysis. Five of the samples were of rocks belonging to the gabbroic suite, one to the transitional suite, four to the dioritic suite and four were of the aplitic and pegmatitic injections. These analyses, listed on Table I (*a*), show an increase in the silica content from the gabbroic type to the aplitic and pegmatitic varieties, and it will be seen that the silica ranges of the different types do not overlap. The Niggli molecular values and the basic molecular norms, as calculated from the analyses, appear in Tables I (*b*) and I (*c*) respectively.

In order to deduce the possible origin and to ascertain the course of crystallisation of the Yzerfontein rocks, several variation curves, based on the norms, the Niggli-values and the relative proportions of the various oxides, were prepared. These were compared with standard curves.

Early in the mineralogical investigation it was felt that, owing to the prominent development of orthoclase, the series may show potassic rather than calc-alkaline affinities. A diagram of the k'-type (Fig. 5), which represents the relative amounts of normative feldspar, shows, however, that except for the aplite and pegmatite analyses (Nos. 11, 13 and 14) the entire series belongs to the calc-alkali suite.

TABLE I (*a*).

Chemical Analyses.

	1.	2.	3.	4.	5.	6.	7.	8.	9.	10.	11.	12.	13.	14.	D.	R.
SiO_2 ...	43·23	49·02	49·84	51·28	51·56	54·28	55·78	55·90	57·26	56·31	61·04	66·60	63·92	71·48	56·31	63·98
Al_2O_3 ...	9·71	13·13	14·92	17·05	15·42	16·06	15·40	15·66	16·00	16·40	18·00	15·60	19·93	12·60	15·87	16·45
Fe_2O_3 ...	6·72	5·60	4·34	4·00	5·11	3·20	4·16	3·04	2·72	3·04	0·80	1·45	0·96	3·19	3·24	1·06
FeO ...	8·33	6·61	6·03	5·03	5·75	4·60	4·02	4·60	4·45	4·02	0·14	2·29	—	tr	4·27	6·05
MgO ...	15·15	7·65	7·66	5·21	5·79	4·89	4·62	4·58	4·04	4·16	0·59	2·12	1·07	0·07	4·35	3·06
CaO ...	9·20	9·62	9·30	8·62	7·80	7·38	5·96	6·22	5·82	6·00	5·32	1·08	2·18	1·62	6·00	1·80
Na_2O ...	1·04	2·50	3·03	2·54	2·81	2·71	2·93	2·84	2·90	3·32	6·47	2·14	8·38	2·14	3·00	3·72
K_2O ...	1·39	1·97	1·67	2·69	1·88	3·88	3·90	3·82	4·46	3·60	3·15	4·25	1·33	6·85	3·95	3·07
$+H_2O$...	2·70	1·37	1·38	1·34	1·63	1·12	1·36	1·47	0·79	1·31	0·64	1·87	0·90	0·44	1·23	—
$-H_2O$...	0·08	0·09	0·06	0·08	0·04	0·03	0·06	0·11	0·08	0·07	0·07	0·11	0·13	0·55	0·08	—
CO_2 ...	—	—	—	—	—	—	—	—	—	—	2·99	—	—	—	—	—
TiO_2 ...	1·08	1·10	0·90	0·93	1·13	0·79	0·80	0·76	0·81	0·79	0·33	0·58	0·76	0·31	0·79	0·55
S ...	—	—	—	0·02	—	tr	tr	tr	tr	tr	tr	tr	tr	1·02	—	—
MnO ...	0·26	0·24	0·18	0·19	0·17	0·19	0·17	0·15	0·15	0·16	0·02	0·09	0·03	0·02	0·16	—
P_2O_5 ...	0·74	0·71	0·48	0·38	0·42	0·43	0·45	0·44	0·43	0·43	0·07	0·21	0·08	0·08	0·43	0·26
Total ...	99·63	99·61	99·79	99·54	99·51	99·52	99·61	99·59	99·91	99·61	99·63	99·55	99·67	99·87	99·68	100·00

TABLE I (*b*).

Molecular Values.

	1.	2.	3.	4.	5.	6.	7.	8.	9.	10.	11.	12.	13.	14.	D.	R.
si ...	82·3	114·1	116·7	132·8	133·1	150·1	162·9	164·1	166·8	168·0	233·3	295·2	247·7	394·4	165·5	231·0
al ...	10·8	17·9	20·5	25·9	23·4	26·0	26·4	27·1	27·4	28·8	40·4	40·7	45·3	41·1	27·4	35·2
fm ...	66·8	49·7	46·9	39·3	45·0	38·1	39·6	38·6	37·8	35·8	6·2	27·7	9·1	13·9	37·9	37·8
c ...	18·7	23·9	23·3	23·9	21·5	21·8	18·6	19·2	18·2	19·1	21·8	5·3	9·1	9·6	18·9	6·9
alk ...	3·7	8·5	9·3	10·9	10·1	14·1	15·4	15·1	16·6	16·3	31·7	26·3	36·5	35·4	15·4	20·2
k ...	0·47	0·34	0·27	0·41	0·31	0·48	0·47	0·47	0·51	0·42	0·25	0·46	0·14	0·68	0·47	0·35
mg ...	0·65	0·54	0·57	0·51	0·50	0·53	0·51	0·53	0·47	0·52	0·56	0·51	0·69	0·05	0·51	0·44
c/fm...	0·28	0·48	0·50	0·61	0·48	0·58	0·47	0·50	0·48	0·54	3·52	0·19	1·00	0·69	0·50	0·18
qz ...	−32·5	−19·9	−20·5	−10·8	−7·3	−6·3	+1·3	+3·7	+0·4	+2·8	+6·5	+85·4	+101·7	+152·8	+2·3	+50·0

TABLE I (*c*).

BASIC MOLECULAR NORMS.

	1.	2.	3.	4.	5.	6.	7.	8.	9.	10.	11.	12.	13.	14.	D.	R.
Kp ...	5·1	7·2	6·1	9·9	6·4	13·9	14·0	13·7	16·2	12·9	11·5	15·8	7·1	25·3	14·7	11·0
Ne ...	5·8	13·7	16·2	14·0	14·4	14·9	16·1	15·7	15·9	18·0	35·2	18·2	43·5	11·8	16·4	20·1
Cal ...	10·8	11·5	13·5	16·6	13·7	12·2	10·8	11·6	10·5	11·9	6·4	2·9	5·8	2·9	11·2	4·3
Cp ...	1·4	1·4	1·1	0·9	0·8	0·9	0·8	0·9	0·8	0·9	0·3	0·3	0·3	0·3	0·9	0·6
Sp ...	—	—	—	—	—	—	—	—	—	—	—	6·2	0·3	—	—	7·2
Cs ...	7·4	7·6	6·3	4·1	14·6	4·3	2·9	2·7	4·5	2·4	4·6	—	—	0·8	3·4	—
Fs ...	7·0	6·0	4·7	4·3	5·1	3·4	4·4	3·2	2·8	3·2	0·8	1·5	1·0	3·5	3·4	1·2
Fa ...	10·7	8·2	7·4	6·2	6·5	5·7	5·0	5·6	5·4	4·9	0·2	2·8	—	—	5·2	7·0
Fo ...	32·5	16·4	16·2	11·1	11·6	10·4	9·9	9·8	8·5	8·8	1·3	1·8	3·3	0·2	9·3	2·8
Ru ...	0·8	0·8	0·6	0·7	0·7	0·6	0·6	0·6	0·6	0·6	0·2	0·5	0·5	0·3	0·6	0·4
Q ...	18·4	27·1	27·9	32·2	26·2	33·7	35·5	36·1	36·0	36·4	39·5	50·1	39·3	55·0	36·0	45·4
L ...	21·7	32·4	35·8	40·5	34·5	41·0	40·9	41·0	42·6	42·8	53·1	36·9	56·4	40·0	41·8	35·4
M ...	57·6	38·2	34·6	25·7	37·8	23·8	22·2	21·3	21·3	19·3	6·9	6·1	3·2	4·5	21·0	11·0
	+2·2	+2·2	+1·7	+1·6	+1·5	+1·5	+1·4	+1·5	+1·4	+1·5	+0·5	+7·0	+1·1	+0·6	+1·5	+7·8
	0·50	0·36	0·38	0·41	0·40	0·30	0·26	0·28	0·25	0·28	0·12	0·08	0·10	0·07	0·27	0·12

1. Coarse-grained olivine(?)-gabbro, North Beach, Yzerfontein.
2. Medium-grained hypersthene-gabbro, "Promontory", Yzerfontein.
3. Coarse-grained leuco-gabbro, North Beach, Yzerfontein.
4. Coarse-grained, poikilitic monzo-gabbro, "Promontory", Yzerfontein.
5. Fine-grained gabbro inclusion in diorite, quarry at harbour, Yzerfontein.
6. Medium-grained, transitional, gabbro-diorite, the Eiland, Yzerfontein.
7. Medium-grained augite-biotite-diorite, south of "Promontory", Yzerfontein.
8. Medium-grained augite-biotite-diorite, bathing beach, Yzerfontein.
9. Medium-grained augite-biotite-diorite, Vlaekop beacon, Yzerfontein.
10. Medium-grained augite-biotite-diorite, quarry at harbour, Yzerfontein.
11. Fine-grained soda-aplite, the Eiland, Yzerfontein.
12. Fine-grained potash-soda-aplite, Gladdeklip, Yzerfontein.
13. Coarse-grained soda-pegmatite, quarry at harbour, Yzerfontein.
14. Coarse-grained potash-soda-pegmatite, Yzerfontein.

D. Average analysis of four Yzerfontein diorite analyses.
R. Composite analysis of two highly contaminated, marginal granites, Darling pluton.

Analyst: C. J. Liebenberg, except analysis R, which is quoted from Scholtz (1946).

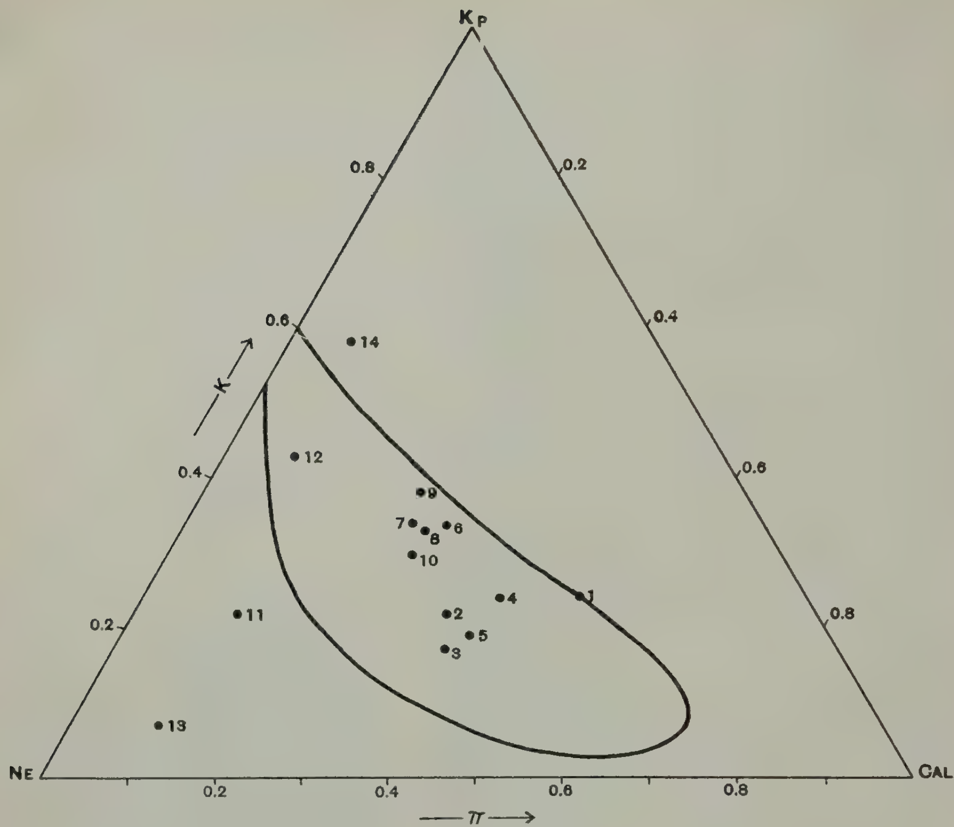


Fig. 5.

In Fig. 6, the various Niggli values have been plotted against the *si*-values as abscissa. The analyses of the aplitic suite have been excluded from this diagram as their abnormal composition is indicated by Fig. 5, and the crystallisation of these rocks will be discussed later, while, because of their similarity, the molecular values of the rocks of the dioritic suite (analysis No. 7, 8, 9 and 10) have been averaged to yield the type D, with an *si*-value of 165·5.

Examination of Fig. 6 discloses that the analyses of the different rock suites are limited to definite *si* ranges, the analyses of the gabbros being spread between $si = 82$ and 133 and the transitional rock having an *si*-value of 150, while the analyses of the diorites are grouped between $si = 163$ and 168. Further, the molecular values of the gabbros and diorites are seen to vary regularly and continuously within their respective suites, but a distinct break in the trends of these curves is apparent between $si = 140$ and $si = 163$, and it is therefore possible to distinguish separate courses of chemical variation in these two suites. It is interesting to note that the molecular values of the transitional rock (analysis No. 6) illustrate the intermediate character of this type by being located within this gap, while its values do not differ markedly from those of a theoretical calc-alkali differentiate of analogous silica content.

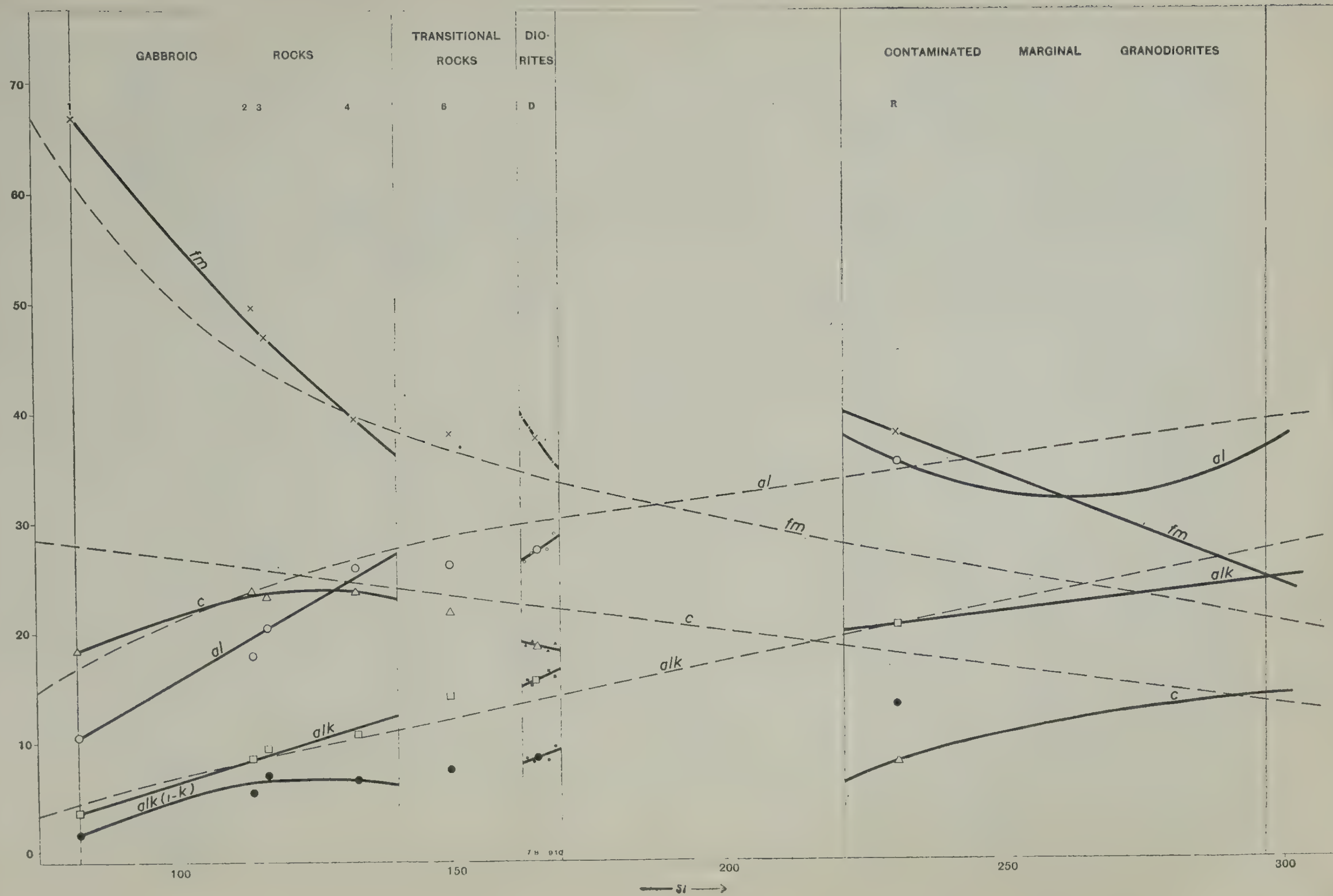

GABBROIC ROCKS
TRANSITIONAL ROCKS
DIORITES
CONTAMINATED MARGINAL GRANODIORITES
1
2 3
4
6
D
R
7 8 9 10
fm
al
c
alk
alk (1-k)
70
60
50
40
30
20
10
0
100
150
200
250
300
si

FIG. 6.

TABLE II.

MODES.

	1A.	2.	3.	3A.	3B.	4.	5.	6.	7.	8.	9.	10.
Quartz	—	—	—	—	—	—	1·8	1·9	4·5	5·0	5·3	3·0
Micropegmatite	—	—	—	—	—	—	—	35·0	33·3	33·0	25·4	42·3
Orthoclase ...	17·3	16·5	11·3	15·8	14·2	34·7	13·7		—	—	—	—
Plagioclase ...	41·8	37·7	44·3	43·9	34·6	28·9	49·5	31·8	32·3	34·7	37·0	27·8
Olivine... ...	—	—	—	—	—	—	—	—	—	—	—	—
Hypersthene ...	12·6	10·0	9·3	8·8	10·0	4·7		6·1	—	—	—	—
Augite	20·0	21·9	20·3	19·1	26·0	19·2	28·3	14·7	8·9	7·5	10·2	8·3
Hornblende ...	—	—	—	—	—	—		tr.	9·5	9·9	11·9	8·8
Biotite	3·6	6·5	8·9	7·4	9·0	6·9	8·3	5·1	6·3	5·7	6·9	5·0
Iron ore ...	3·8	5·9	4·8	4·1	4·9	4·5	7·6	4·4	4·3	3·0	2·4	4·0
Apatite... ...	0·9	1·5	1·1	0·9	1·3	1·1	0·8	1·0	1·0	1·2	0·9	0·8
Total ...	100·0	100·0	100·0	100·0	100·0	100·0	100·0	100·0	100·0	100·0	100·0	100·0

A trend in common with the variation in all the suites is the rising of the *alk*-value with increasing acidity and this may be expected to reflect the increase of the proportion of alkali feldspar as crystallisation proceeded. In the gabbros this augmentation is not equally shared by sodic and potassic feldspar as may be shown by the *alk*(l-*k*) curve, which indicates the proportional development of normative albite. Here the initial increase of albite changes gradually to a slight reduction, while normative orthoclase shows continual enrichment.

The *al*-curve also exhibits a continuous upward trend in the gabbros, but with a slope steeper than that of the *alk*-curve, indicating by the difference (*al-alk*) an increasing normative anorthite. Mineralogically, however, a slight decrease in the amount of this mineral is observed and the explanation of this phenomenon probably lies in the fact that both the biotite and the augite contain aluminium. It is significant that the modal proportions of the two last named minerals are low in the most basic gabbros and increase with increasing acidity. Biotite retains this trend throughout the series while a decrease in the proportion of augite becomes apparent in the most acid gabbros.

When the behaviour of augite is considered in conjunction with the *c*-curve, the decreasing anorthite content of the plagioclase may be further demonstrated. If

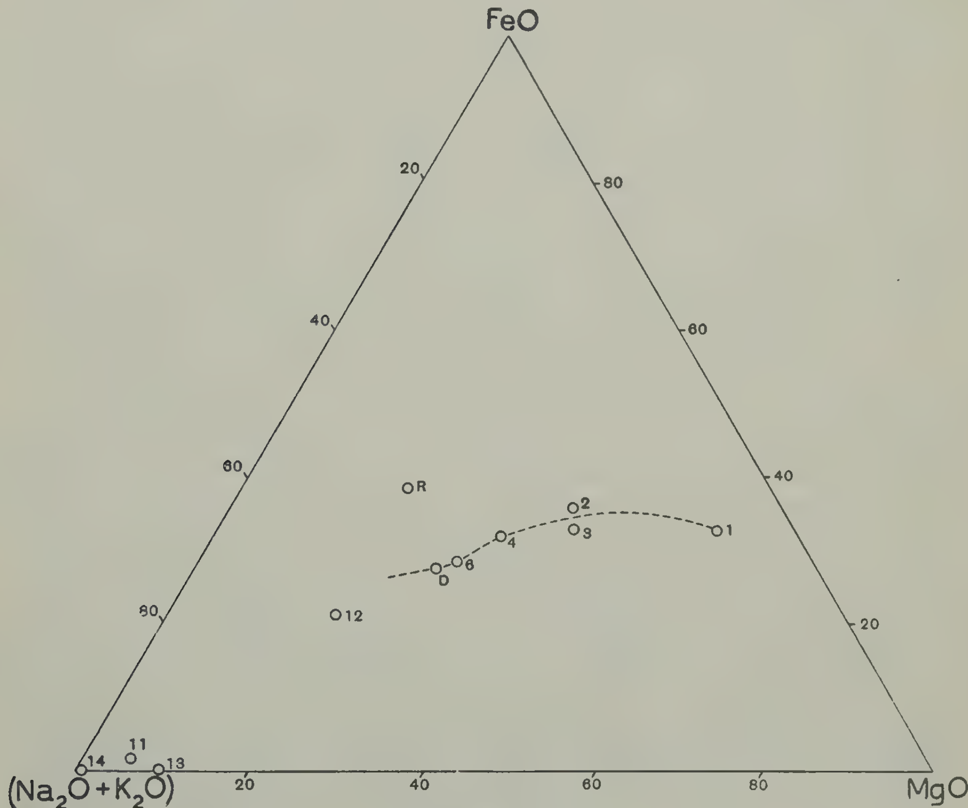


FIG. 7.

augite is disregarded, then the value of *c-alk*(1-*k*) will represent the anorthite content of the plagioclase, and since the *c*- and *alk*(1-*k*)-curves are essentially parallel, this anorthite content would remain constant throughout the suite. The addition of augite would now decrease this value proportionally. Thus, increasing augite with increasing acidity must result in a decrease of the anorthite content of the plagioclase.

The behaviour of the other dark minerals may be demonstrated by the *fm*-, *mg*- and *c/fm*-values. The steep downward trend of the *fm*-curve designates a pronounced decrease of the mafic minerals with decreasing basicity, while the distinct increase of the ratio *c/fm* from 0·28 to 0·61 shows that this decrease is probably entirely due to the rapid diminution in the amount of olivine and orthopyroxene. This trend is further accompanied by a decrease of the *mg*-value from 0·65 to 0·51 which may reflect the early crystallisation and separation of olivine.

Many of these features may be verified by the trends of variation illustrated in Figs. 7 and 8.

In Fig. 7 the relative proportions of the alkalis, ferrous iron and magnesia have been plotted on a ternary diagram. In the gabbroic suite, it is seen that the initial increase of iron and the alkali content and the decrease of magnesia is followed by pronounced alkali enrichment with little variation of the FeO/MgO-ratio.

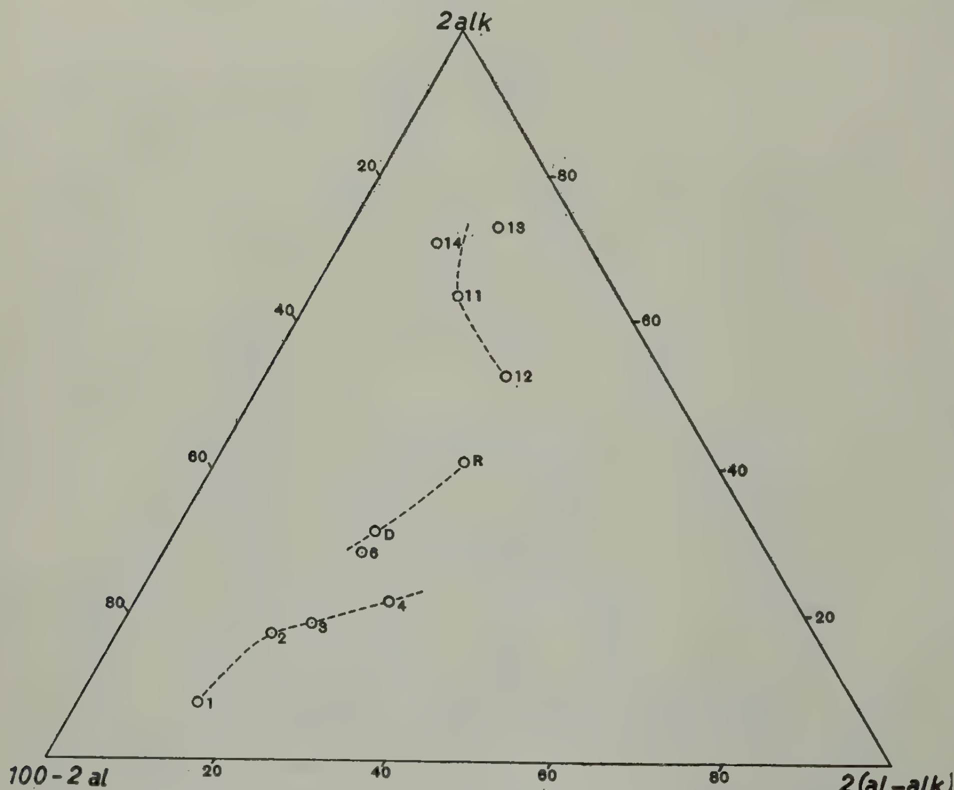


FIG. 8.

Fig. 8 is a ternary plot of (100-2*al*), 2*alk* and 2(*al-alk*), which represent respectively the normative proportions of the dark minerals, alkali feldspar and anorthite. Here the pronounced decrease of dark minerals coincides with a general increase of feldspar, initially alkali- and subsequently lime-enrichment. Van Zyl (1950) shows that both these trends characterise the course of crystallisation of the Newry suite of igneous rocks.

The passage from the gabbro to the diorite across the transitional zone involves an *si*-increase of approximately 20 units, the change being demonstrated across the relevant gap in Fig. 6. The alkali enrichment which accompanies this transition is due almost entirely to an increase in soda feldspar, potash remaining essentially constant as shown by the sympathetic relationship between *alk* and *alk*(1-*k*). A pronounced reduction of anorthite is indicated by the decrease of the *al*-value and is corroborated by the similar behaviour of the *c*-value. On the other hand, increasing *fm* shows an addition of mafic minerals. These changes also appear in Figs. 7 and 8, the latter showing clearly that the transition from the gabbro to the diorite is accompanied by pronounced alkali enrichment, a slight increase of the mafic constituents and a notable reduction in the anorthite content.

In the dioritic suite, the trend of variation apparently parallels that of the gabbroic suite, and here too the increase of normative anorthite may be explained by the presence of such aluminium-bearing minerals as amphibole, biotite and augite. The further downward trend of the *c*-curve indicates an actual reduction of both anorthite and diopsidic augite.

For comparative purposes the standard curves of Niggli's calc-alkaline suite of rocks have been incorporated in Fig. 6. While the molecular values of the most basic gabbros and diorites differ markedly from the standard values, the trend of the variation in these suites results in the approaching and coincidence of their variation curves with the corresponding standard curves. This is particularly well displayed by the *fm*- and *al*-curves. The convergence is probably due to the fact that the minerals of early crystallisation were composed of those constituents which, in the original magma, exceeded their normal proportions, and this resulted in a concentration of the remaining constituents in the rest magma. Such a trend of variation is believed to be due to a process of normal and undisturbed differentiation.

Among the analyses of Yzerfontein rocks which were not used in the preparation of the variation diagrams, analysis No. 5 (Table I) is of a representative sample of the dark, fine-grained, rounded inclusions collected at the quarry. In hand specimen this rock has the appearance of an even-grained hornstone, but in thin section it possesses a texture which may be described as microporphyritic, occasional phenocrysts of pyroxene and plagioclase being scattered in an aphanitic groundmass. The freshest samples, such as are found at the cores of the largest inclusions, are mineralogically similar to the gabbros, the phenocrysts being composed of unresorbed crystals of augite and labradorite, while the groundmass contains the same minerals and is moreover usually free from quartz, micropegmatite and hornblende. Towards the margins of the inclusions, however, quartz becomes more prominent and the augite grains have frequently been altered to green hornblende. The texture and mineralogy suggest that this rock was originally the chilled phase of the gabbro magma, and the alteration which it often displays may be attributed to its envelopment by the diorite, a position in which the selective transfer of material between the solid xenolith and the liquid diorite must have proceeded. Chemically, the analysis (No. 5, Table I) also shows an affinity to the gabbroic suite, while its very slight departure from the normal gabbro variation curves is possibly the effect of soaking in the composite dioritic liquid.

The various aplitic and pegmatitic injections, on the other hand, show affinity to a granitic suite and probably represent the pegmatitic and aplitic phases of the adjacent Darling granite pluton. Analysis No. 11 and 12 show that two distinct types of aplite occur at Yzerfontein:—

(i) a soda-aplite with $k = 0 \cdot 25$ and

(ii) a potash-soda-aplite with $k = 0 \cdot 46$.

Both have comparable fine grain and pink colour and are impossible to distinguish in the field. Although they both invade and are younger than all the other igneous rocks at Yzerfontein, it was not found possible to determine their relative ages. The soda-aplite seems to occur mainly in the form of more regular veins and extends further into the gabbro, while the more abundant potash-soda-aplite, though usually comprising the irregular pocket-like bodies in the diorite, may also occur as more regular vein fillings.

The soda-aplite is composed almost entirely of a fine-grained groundmass of albite and subordinate quartz with occasional microcline or albite phenocrysts. Much calcite, epidote and some pyrite represent later additions, while fine hematite dust gives the rock its pink colour. The potash-soda-aplite has a more varied mineralogical composition, microcline, albite, quartz, chlorite, amphibole and accessory blue tourmaline occurring as well as the later minerals.

Pegmatite dykes are not well developed at Yzerfontein, but the frequent occurrence of irregular, lenticular and stringer-like patches of coarse-grained, leucocratic material was taken to represent pegmatite injections. Here too, soda-pegmatites and soda-potash-pegmatites, corresponding to the two aplite types, may be distinguished on both mineralogical and chemical grounds (Table I, Nos. 13 and 14).

In his investigation of the pre-Cape Granite plutons of the George district, C. T. Potgieter (1950, pp. 378-383) also records the presence of two different pegmatites; a quartz-microcline-plagioclase pegmatite, and a quartz-plagioclase pegmatite. He discusses in detail the two main theories accounting for the formation of mineralogically distinct but consanguineous pegmatites, independently emplaced and derived from immiscible potash- and soda-rich residual magmatic products, as theoretically envisaged by S. J. Shand (1948). This hypothesis appears to account most satisfactorily for the presence of both soda- and potash-soda-aplites and pegmatites at Yzerfontein.

IX. AGE RELATIONS

The basic rocks of Yzerfontein appear to be situated along the western margin of the Darling pre-Cape granite pluton, but owing to the absence of any granite outcrops in the immediate vicinity of these rocks, their mutual relationships could not be determined in the field. The sequence could, however, be deduced by the evidence supplied by the aplite, pegmatite and hydrothermal veins. It seems likely that these injections comprise material related to the late magmatic period of the adjacent granites, and that the fissures and joint planes which they have exploited in the diorites all follow directions parallel to those characteristically developed in the granites during their final stages of consolidation. The fact that these aplite, pegmatite and hydrothermal mineral-bearing veins and dykes traverse and include xenoliths of both the diorites and the gabbros suggests that the rocks were emplaced contemporaneously with, or more probably earlier than, the intrusion of the granite. This is further borne out by the shearing and mylonitisation of the diorites by forces which produced analogous and parallel cataclastic structures in the older pre-Cape granites.

It has already been demonstrated that, within the basic complex itself, intrusions of two different ages are present, namely, an earlier gabbroic and a later dioritic intrusion holding xenoliths of the gabbro. Assigning to the diorite its youngest possible age, that is, synchronous with the intrusion of the granite magma, the pre-granite age of the gabbro magma is definitely indicated.

X. PETROGENESIS

(i) The Gabbro.

It is believed that the form of the early intrusion of basaltic magma was either a laccolith or a sheet-like body characterised by a sub-horizontal floor. Evidently chilling of the magma occurred where it came into contact with the colder rocks forming the roof and floor of the intrusion, but, although rounded xenoliths of the fine-grained chill phase are frequently encountered in the later diorites, the actual contacts are nowhere exposed.

During the period of quiescene following on the intrusion of the magma, fractional crystallisation proceeded apace and the early precipitated crystals of olivine, pyroxene and basic plagioclase settled to the base of the intrusion under the influence of gravity. Owing to the differential crystal settling, a well-defined primary banding and igneous lamination developed parallel to the floor of the intrusion. The removal of these early crystals left a magma enriched in potash and alumina, some of which was trapped in the interstices between the settling crystals of the banded gabbro, where it finally crystallised as biotite and orthoclase. However, the greater part of the rest-magma concentrated towards the top of the body and here, as has been demonstrated by J. H. L. Vogt (1923, pp. 233-239), the progressive enrichment in potash and alumina as crystal fractionation proceeded must have greatly reduced its fluidity. Under conditions of increasing viscosity, the settling of the early-formed crystals was arrested and orthoclase, which crystallised from the potash-enriched residual liquid, enveloped them in a poikilitic fashion to form the mottled monzonitic facies of the gabbro.

The course of crystallisation of the gabbro suite, as reflected by the variation diagrams, follows that displayed by the Newry suite of igneous rocks, a trend which P. J. van Zyl (1950) erroneously generalises as characterising the contamination of an acid magma by basic material, and the presence in this region of material more basic than the gabbro could not be entertained. When compared with the variation of the molecular values of Niggli's standard calc-alkali suite, however, the variation of the Yzerfontein gabbros may be explained as a product of normal differentiation by crystal fractionation along a gravity gradient.

The existence in the south-western Cape Province of a pre-granite basaltic magma was first suggested by P. J. van Zyl (1950) who found that this magma had been emplaced into a dome-structure in the pre-Cape sediments along the south-western margin of the Malmesbury-Paardeberg granite pluton after it had undergone slight modification by differentiation in depth. It is interesting to note that both the Malmesbury and Yzerfontein gabbro suites possess relatively high *fm*-values and rather low *c*-values, but appear to have followed individual, dissimilar differentiation trends, due no doubt to the different physical conditions under which the magma was emplaced. At Malmesbury differentiation occurred in depth and the variation observed is due to successive intrusions with little differentiation *in situ*, while at Yzerfontein all

differentiation may be ascribed to processes succeeding the emplacement of the magma.

The Malmesbury diorite intrusions produced intense contact metamorphic alteration of the country rock and the secondary structures of these metamorphosed sediments appear to be wrapped around the diorite stocks. Owing to the absence of outcrops of pre-Cape sediments in the vicinity of Yzerfontein, it was unfortunately impossible to deduce either the relationship which might exist between the attitude of the gabbro body and the structure of the surrounding, invaded sediments, or to observe the metamorphic effect the former had had on the latter. It is, however, noteworthy that the basic intrusions under consideration are located along the south-western flanks of the Malmesbury and Darling pre-Cape granite batholiths, that is, on the side from which the pressure, which caused the elongation of these last-mentioned bodies, operated during their emplacement. No other gabbroic or dioritic bodies of analogous form or equivalent size are known to occur on the north-eastern flanks of the granite plutons.

(ii) **The Diorite.**

Any discussion bearing on the probable genesis of the Yzerfontein diorites must take into account the four main hypotheses which have been advanced for the origin of this rock clan in general. Thus, diorite may be the product of:—

(*a*) the normal differentiation of a basic magma,
(*b*) the reaction between a basic magma and acid country rock,
(*c*) the reaction between acid magma and limestone,
(*d*) the reaction between acid magma and basic igneous rock.

The applicability of each of the above-mentioned hypotheses in the case of the Yzerfontein diorites may be critically reviewed.

(*a*)

The discordant attitude of the dioritic suite towards the primary structure of the gabbro intrusion precludes the possibility that these rocks have resulted from the *in situ* differentiation of the earlier gabbro magma. However, by a process similar to that which led to the intrusion of diorite at Malmesbury, the Yzerfontein diorite may be the product of a subsequent heave of the basaltic magma which had undergone differentiation in depth. It has, however, already been pointed out that the trends of the variation curves of the Malmesbury and Yzerfontein rocks are markedly dissimilar. Moreover, petrographic studies disclose the presence of a considerable amount of gabbroic material in the Yzerfontein diorites and this necessarily implies that the composition of the second intrusion must have been decidedly more acid than that of the resultant dioritic rock. Additional evidence of the more acid nature of this magma is afforded by the extensive reaction which had apparently taken place between it and the gabbroic xenocrysts, but there is no field evidence to justify the former existence of a large body of basaltic magma capable of yielding adequate quantities of an acid differentiate.

(*b*)

Any theory of the origin of the diorites which involves the contamination of the gabbroic magma by acid country rock would also be unable to explain the discordant transgression of the gabbro by the diorite. That the latter was derived from a later fluid intrusion is further indicated by the transitional zone where a plastic mixing of the two types is clearly visible. The total absence of peraluminous minerals or sedimentary xenoliths in the diorite also militates against any such hypothesis.

(*c*)

Further to the east, occasional thin and rather impure limestone beds are known to be present in the pre-Cape formations which have been invaded by granite. Small outcrops of metamorphosed calcareous rocks occur near the southern extremity of the Darling pluton, while the Robertson boss cuts across thin limestone beds, typical metamorphic lime silicates being developed along the contacts. No diorites are, however, known in these areas. (Scholtz, 1946, pp. liv.)

The pre-Cape rocks in the Yzerfontein-Saldanha region are apparently devoid of primary limestone beds and, since the diorites under consideration provide no positive mineralogical or chemical evidence of the addition of lime, any hypothesis postulating the contamination of granite by calcareous rock may be dismissed without further comment.

(*d*)

It has been previously suggested that the diorite could have resulted from the incorporation of gabbroic material in a later magma considerably more siliceous than the resultant diorite (i.e. si $> 165 \cdot 5$). The existence of such a magma in this vicinity is demonstrated by the granitic rocks of the Darling pluton, and the fact that Yzerfontein is situated close to the margin of this batholith may be deduced from the profusion of aplites, pegmatites and other marginal veins and dykes related to the final stages of the consolidation of the granite. D. L. Scholtz has shown that this pluton has developed a highly contaminated marginal facies by the incorporation and assimilation of shaly country rock, and it is believed that this contaminated magma, rather than the more normal granite magma, reacted with the gabbro to produce the diorite. This is verified by the almost perfect linear relation between the compositions of the marginal granite (analysis R, Table I), the average diorite (analysis D) and the gabbro (analysis No. 4), which permits a quantitative synthesis of the diorite. To quote a particular example, we may select the most acid gabbro analysis (Table I, No. 4) and add to this such a proportion of the marginal granodiorite (analysis R) that on recalculation on the basic molecular values to 100, the *si*-value of this mixture will correspond to that of the average Yzerfontein diorite (analysis D). Thus, if two parts of gabbro analysis No. 4 are added to the average marginal granodiorite analysis R, a hybrid with *si*-value of $165 \cdot 7$, corresponding to $si = 165 \cdot 5$ of the average diorite analysis D, is produced. A remarkable agreement between the molecular values of the calculated and the average Yzerfontein diorite can be demonstrated:—

	si	*al*	*fm*	*c*	*alk*
R	231	35·1	37·8	6·9	20·2
2 × no. 4	266	51·8	78·6	47·8	21·8
Sum	497	86·9	116·4	54·7	43·0
∴ Calculated diorite	165·7	29·0	38·8	18·2	14·3
Average diorite (D)	165·5	27·4	37·9	18·9	15·4

Considering the variable composition of both the gabbro and the contaminated granodiorite, the correspondence of the basic molecular values of the calculated and actual diorite is exceptional, but still better agreement is shown when $63 \cdot 7\%$ of gabbro analysis No. 4 is added to $36 \cdot 2\%$ of a slightly more basic granodiorite from the Darling pluton (analysis No. 42 of Scholtz (1946)).

	si	*al*	*fm*	*c*	*alk*
No. 42	223	33·5	37·7	6·4	22·3
36·2% of no. 42 + 63·7% of no. 4	165·5	28·7	38·7	17·7	15·0
Average diorite (D)	165·5	27·4	37·9	18·9	15·4

Like results may be obtained from similar calculations in which various contaminated granodiorite analyses are added to interpolated gabbro analyses read off from the more acid portion of the gabbro differentiation curves in Fig. 6.

The calculations all demand, however, that a given mass of granodiorite should incorporate approximately twice as much gabbro, and the source of the heat necessary becomes problematic. Volumetric comparisons show that the amount of diorite formed is but a fraction of the volume of the granite intruded, and it is possible that a considerable supply of heat was available from the centre of this mass. According to D. L. Scholtz, the upward transfer of heat by convection currents during the emplacement of the Darling pluton appears to have been more effective in this batholith than in most of the other younger pre-Cambrian granite plutons of the Cape Province. Moreover the available petrographic evidence suggests that the granite magma invaded the gabbro body prior to its complete consolidation. In this connection, it may be recalled that, instead of a sharp, intrusive contact displaying evidence of chilling of the granite against the gabbro, a broad transitional zone exists in which the plastic mixing of the two types is apparent—a relation which implies that the heated gabbro was, in all probability, in a semi-plastic or crystal-mush state when the later acid magma was discordantly emplaced from the east. This conclusion is also substantiated by the rarity of gabbroic inclusions in the diorite, other than that of the fine-grained chilled marginal variety, which seems to indicate that the chill phase of the gabbro had already congealed and was sufficiently cold and compact to effectively resist complete digestion by the younger, intrusive, contaminated pre-Cambrian granite.

The attitude of the primary banding of the gabbro suggests that originally the greatest portion of this body was probably situated some distance below the present diorite outcrops. Although some of the gabbro had undoubtedly consolidated before the intrusion of the granite, most of the body was presumably still at a relatively high temperature and the more acid differentiate was probably mobile although fairly viscous. The granite magma, its composition already modified by the assimilation of argillaceous country rock, apparently invaded the gabbro body from below and was further altered by the addition of the gabbroic rest magma and by the incorporation of already crystalline gabbroic material in the form of both single crystals and semi-plastic crystal aggregates. Although considerably more basic than the original granite magma, this contaminated liquid was still sufficiently acid to react with the foreign solid phases, earlier in the reaction series, with the production of coronas about discrete crystals. Disruption of the larger mushy inclusions could have been effected by active mechanical disintegration, by passive chemical reaction, or by a combination of these processes. The hybrid magma so formed rose to a higher level where, by further mixing with gabbro in the act of crystallisation, it gave rise to the rocks of intermediate composition characterising the transitional zone.

In most examples of the hybridisation of an acid magma by the assimilation of basic material the final product is almost invariably heterogeneous in character. Both W. A. Deer (1935) and G. D. Nicholls (1951) have, however, observed that under

favourable conditions, the product of hybridisation may be completely uniform rock having the appearance of a normal igneous type which reveals little or no evidence of its hybrid origin. On the one hand, in the Cairnsmore of Carsphairn Igneous Complex such conditions apparently obtained and here the hybridisation of granitic magma by partly solidified gabbro took place in depth before the intrusion attained its present position. The resultant tonalitic hybrid is completely uniform (Deer, 1935, p. 66). On the other hand the petrological uniformity of the Eastern Adamellites of the Glenelg-Ratagain Igneous Complex is ascribed to the fact that the assimilated material had a composition similar to the phase in equilibrium with the intruded magma (Nicholls, 1951, p. 338). The gabbroic minerals in the Yzerfontein diorites, however, show evidence of extensive reaction with the later magma and it is unlikely that the last named mechanism of hybridisation could account for the homogeneity of these rocks. On the contrary, the history of formation of the Yzerfontein diorites and the Carsphairn tonalites is remarkably similar, and the comparative uniformity of the diorites may be ascribed to hybridisation and homogenisation at depth prior to intrusion. It is interesting to note that S. R. Nockolds (1933, p. 589) suggests that the movement of a magma is a potential aid to the mechanical disintegration of included xenoliths, and the products of such activity are either small, rounded clots of foreign material or individual xenocrysts. Moreover, the motion tends to distribute the solid phases set free from the xenoliths in an even manner throughout the contaminated magma, thus obscuring its hybrid nature. Whereas the homogeneity of the Yzerfontein diorites, therefore, suggests movement after hybridisation of the granodiorite magma, the heterogeneity of the transitional hybrid is suggestive of origin *in situ*.

Another common feature of diorites which have been formed by the reaction between an acid magma and basic rocks is the prominence of the role of reciprocal reaction. By this process, which involves the selective transfer of certain constituents from the magma to the basic rocks and vice versa, the reaction products, whether contaminated magma or reconstitued country rock, are rarely simple addition products of the pure magma and basic rock. As the Yzerfontein diorite does show such linear relationship between the gabbro and granodiorite, the role of reciprocal reaction appears to have been negligible. In general, linear variation has been explained by:

- (i) the mixing in all proportions of two different liquid magmas (Wilcox, 1944);
- (ii) the completion of the reciprocal reaction process under exceptionally favourable conditions (Thomas and Smith, 1932);
- (iii) the mechanical incorporation in the magma of assimilated material *en masse* (Deer, 1935; Nicholls, 1951).

At Yzerfontein all three processes have probably been active in varying degrees, but only the unselective incorporation of gabbroic material has occurred on a sufficiently large scale to produce the linear variation observed. This essentially mechanical process which, as we have already seen, was also responsible for the uniformity of the diorites, was probably greatly assisted by the pre-heated and semi-plastic nature of much of the gabbro.

From the foregoing petrological investigation, the writer is of the opinion that the following sequence of events probably occurred in the igneous history of the Yzerfontein Complex:—

- (i) The intrusion of a sub-horizontal sill-like or laccolithic body of a basaltic magma, which underwent differentiation *in situ* by crystal fractionation along a gravity gradient. The gabbroic rocks so formed are evidently co-magmatic with the Malmesbury gabbro-diorite suite.

(ii) The emplacement of the younger pre-Cambrian granite of the Darling batholith and the assimilation of argillaceous country rock giving rise to the marginal granodioritic phase of this intrusion.

(iii) Contamination of the granodiorite magma by gabbroic material at Yzerfontein led to the development of a hybrid dioritic liquid. Upward movement of this hybrid liquid induced greater uniformity, while further mixing with uncongealed gabbro resulted in the production of the heterogeneous hybrids of the transitional zone.

(iv) Shortly after their consolidation, these rocks were invaded by late aplitic and pegmatitic differentiates of the Darling granite batholith, primarily along N.W.-S.E. directions of jointing and shearing.

(v) The infilling of some of these joints by later, hydrous, magmatic fluids from which crystallised the various assemblages of hydrothermal minerals was probably connected with the final stages of magmatic activity of the granites.

(vi) The development of S.W.-N.E. orientated joints.

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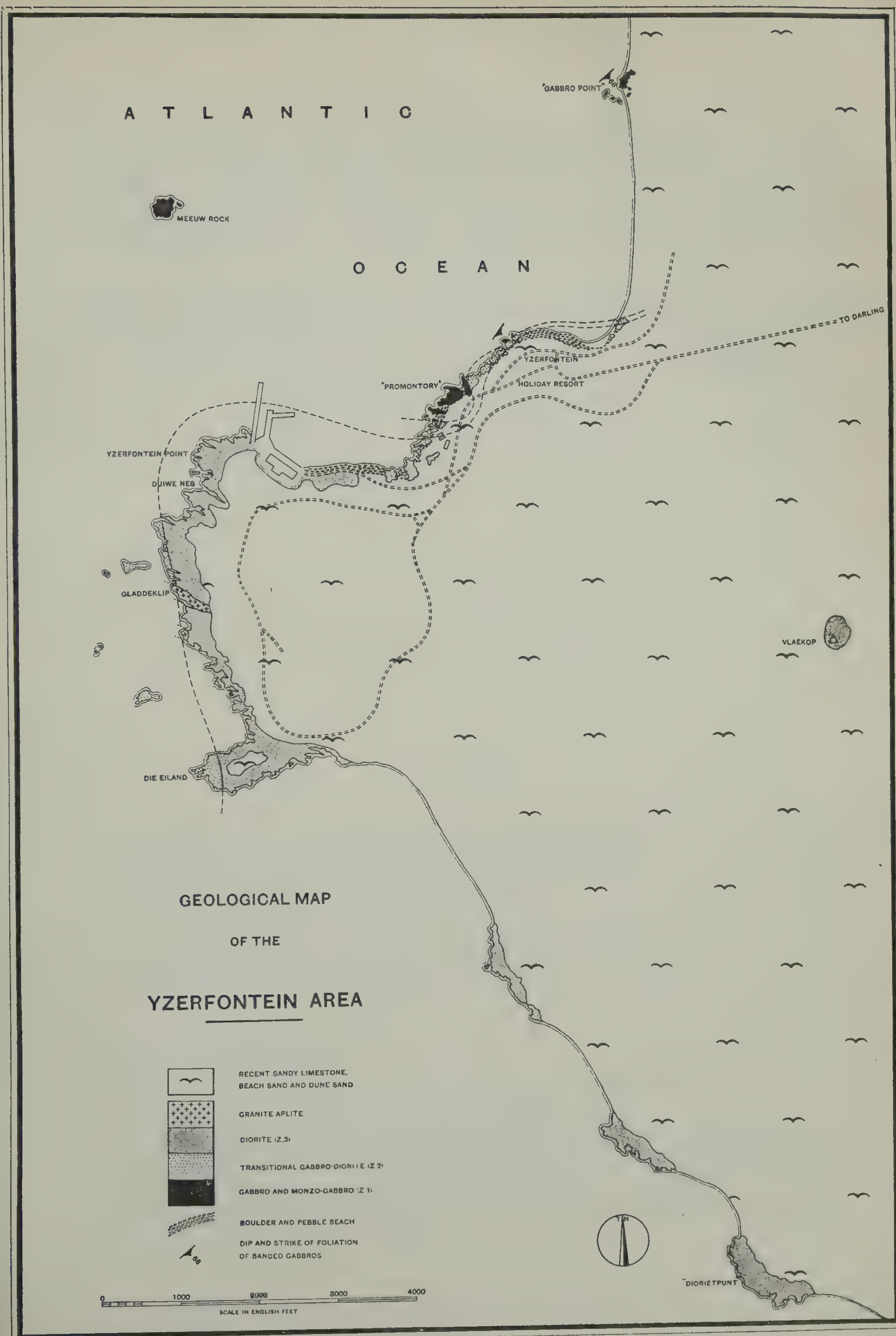

ATLANTIC
OCEAN
GABBRO POINT
MEEUW ROCK
TO DARLING
YZERFONTEIN
HOLIDAY RESORT
PROMONTORY
YZERFONTEIN POINT
DUIWE NES
GLADDEKLIP
VLAEKOP
DIE EILAND
GEOLOGICAL MAP
OF THE
YZERFONTEIN AREA
RECENT SANDY LIMESTONE, BEACH SAND AND DUNE SAND
GRANITE APLITE
DIORITE (Z.3)
TRANSITIONAL GABBRO-DIORITE (Z 2)
GABBRO AND MONZO-GABBRO (Z 1)
BOULDER AND PEBBLE BEACH
DIP AND STRIKE OF FOLIATION OF BANDED GABBROS
0
1000
2000
3000
4000
SCALE IN ENGLISH FEET
DIORIETPUNT
TN

PLATE III

ON THE TECTONIC HISTORY AND SOME RELATED SEDIMENTATIONAL ASPECTS OF THE WITWATERSRAND AND VENTERSDORP SYSTEMS IN THE FAR EAST RAND, DURING UPPER-WITWATERSRAND, PRE-TRANSVAAL SYSTEM TIME.

By

A. F. CLUVER, M.Sc.

(Submitted December, 1951)

ABSTRACT

The Far East Rand is the north-eastern portion of the large basin of the known Upper Witwatersrand System. A contour map of the Main Reef, and to a lesser degree the surface geology, indicate two distinct structural features, namely, a basin and major folds.

The sediments of the Witwatersrand System were deposited on plane surfaces and the surface on which the System was instituted was a plane surface. The occasional unconformities and disconformities within the system are of relatively small degree.

The relative stability of the Johannesburg-Pretoria Old Granite mass is discussed, also its situation in relation to the depositional basin of the Witwatersrand System.

The relative stability of the Old Granite mass to the east of Nigel is discussed.

The attitudes of the sediments of the Central Rand and Far East Rand are due to movements subsequent to the deposition of the sediments.

The thickness of the sediments is related to the degree of tectonic basining of the Witwatersrand System.

The lavas and intrusive rocks contemporaneous with the sediments of the Witwatersrand System are basic in composition.

The significance of the basic and acid phases of the Ventersdorp System is discussed.

The significance of two distinct phases of basic extrusives of the Ventersdorp System, unconformable to each other, in the Odendaalsrus area is discussed.

The probable cause of the tectonic basining of the Witwatersrand System is discussed.

The folds of the Far East Rand and the significance of their form, trend and pitch are described.

The folds are post-Ventersdorp, pre-Transvaal System in age and were probably formed after the basining of the Witwatersrand and Ventersdorp Systems.

The cause of the folds is discussed: there is evidence that the major movement on the Sugarbush Fault was horizontal and occurred in pre-Transvaal System time and that the forces causing this movement were partly responsible for the folds.

CONTENTS

I INTRODUCTION

Location

The area dealt with in this discussion lies in the province of the Transvaal, South Africa, between latitudes 26° 26′ and 26° 6′ and longitudes 28° 35′ and 28° 17′, is commonly referred to as the Far East Rand or East Rand Basin, and forms part of the well-known Witwatersrand Gold Fields. Approximately 320 square miles of this area are underlain by the Main Reef, sometimes called the Main Reef Leader (Carleton Jones, 1936), which is a gold-bearing quartzitic conglomerate seldom more and usually less than four feet thick. In this area the reef has been mined, developed and prospected to a large degree and its structure can be examined in detail.

Historical Review of "Reef" Correlation

Gold was found on the Witwatersrand in the year 1884 and the Main Reef was discovered in 1888. The person accepted as being the discoverer of this reef is George Harrison (Krause, 1946). The name Johannesburg was given to the mining village in the year 1886 (Jeppe).

Numerous prospectors arrived on the Witwatersrand during 1886 and it was during this year that auriferous conglomerates were found on the Far East Rand. By the year 1888 gold was being produced from the Far East Rand by at least two mines, namely, New Modderfontein in the north and Nigel in the south. At this time it was not known that the reefs worked by these mines were one and the same, and it was only in 1904 that Hatch recognised that the reefs were in the same series of the Witwatersrand System, although in 1895 he suggested that this was probably the case and also that the so-called Van Ryn Reef, Nigel Reef, Geduld Reef, etc., were the same conglomerate band. Hatch also came to the conclusion that this was the same as the principal reef mined on the Central Rand, namely, the Main Reef. Even after Hatch's report the reef on the Far East Rand was called by a variety of names. In 1915 Mellor suggested the name Main Reef Leader as being the probable correlation of the Far East Rand reef with the Central Rand Reef.

In 1936 Carleton Jones suggested the name Main Reef and this is the name generally used on the Far East Rand to-day, although there are a number of mines that still call it the Main Reef Leader. In this discussion the principal reef of the Far East Rand will be called the Main Reef.

Underground workings on the Far East Rand have proved that the principal reefs mined are one and the same reef, and it is possible to walk underground on the horizon of this Main Reef from north to south and east to west over the whole of the Far East Rand.

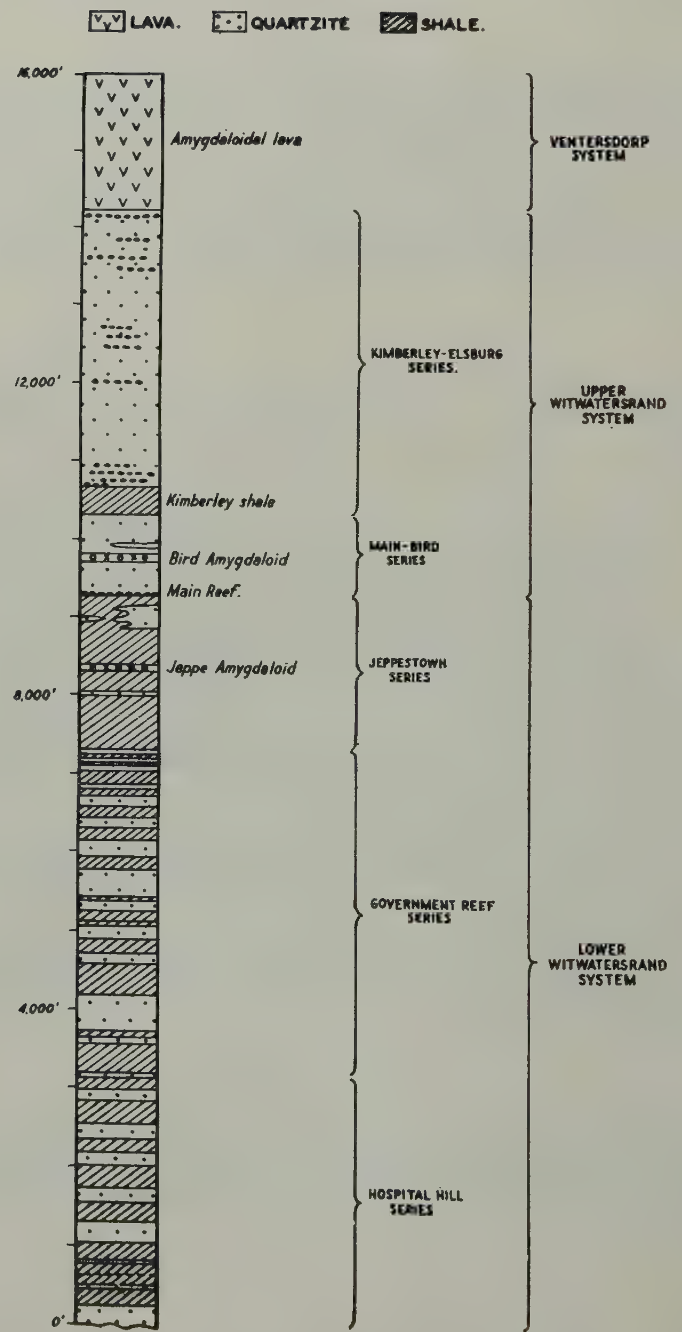
Mining Companies

The area has been and is being mined and prospected by the following mining companies: Van Dyk Consolidated Mines, Limited; New Kleinfontein Company Ltd.; Van Ryn Gold Mines Estate, Ltd.; Modderfontein B Gold Mines, Ltd.;

IDEAL STRATIGRAPHIC SECTION

— FAR EAST RAND —

(AFTER ROGERS)



Modderfontein East, Ltd.; Holfontein (T.C.L.) Gold Mining Company, Ltd.; Welgedacht Exploration Company, Ltd.; Geduld Proprietary Mines, Ltd.; East Geduld Mines, Ltd.; Grootvlei Proprietary Mines, Ltd.; Government Gold Mining Areas (Modderfontein) Consolidated Ltd.; New State Areas, Ltd.; Brakpan Mines, Ltd.; The South African Land and Exploration Company, Ltd.; West Springs, Ltd.; Springs Mines, Ltd.; East Daggafontein Mines, Ltd.; Marievale Consolidated Mines, Ltd.; Vogelstruisbult G.M. Areas, Ltd.; The Sub Nigel, Ltd.; Vlakfontein G.M. Company, Ltd.; West Vlakfontein G.M. Company, Ltd.; Spaarwater G.M. Company, Ltd.; The Nigel G.M. Company, Ltd.; West Spaarwater, Ltd.; The Withok Proprietary Company, Ltd.

Topography

The surface relief is flat to rolling with a few shallow valleys, and the major portion of the area lies 5,350 feet above sea level, the extremes in elevation being 5,543 feet and 5,146 feet above sea level.

The only stream of any importance is the Blesbok Spruit, which, with its tributaries, drains 90 per cent of the area, the drainage being to the south. The surface of the area has been mapped in detail and these maps include contours at 50 foot intervals. Maps of individual mines have surface contours in greater detail. Most of the surface consists of sediments of the Karroo System which overlie the Transvaal, Ventersdorp and Witwatersrand Systems, so that only in isolated patches are the sediments of the Witwatersrand System exposed, and these invariably lie at slightly higher elevations than the surrounding country.

Surface Geology

The surface geology of the area has been described and mapped by E. T. Mellor (1915) and A. W. Rogers (1921) of the Geological Survey of South Africa. Various portions of the area were described previous to these reports, but any relevant facts have been incorporated in the reports of Mellor and Rogers, and for details of the surface geology reference should be made to these reports. Fig. 1 gives an ideal stratigraphic column of the Witwatersrand System on the Far East Rand. The following is a summary of the surface geology of the Far East Rand.

The whole area is underlain by sediments of the Witwatersrand System which, however, outcrops only in the north-western and southern portions of the area under discussion. In the north-western portion only small patches of both the Upper and Lower Witwatersrand Systems are exposed. In this sector the Upper Witwatersrand has been folded and dips in directions varying from south to east. The surface geology indicates that the Upper Witwatersrand has not been disturbed by major faults but that the Lower Witwatersrand is crossed both from east to west and from north to south by faults which have considerable displacement. On the farm Vlakfontein 18 one of these north-south faults has a displacement of some thousands of feet downwards on its east side in the beds of the Lower Witwatersrand, but it has very little displacement in the Upper Witwatersrand (see Fig. 2).

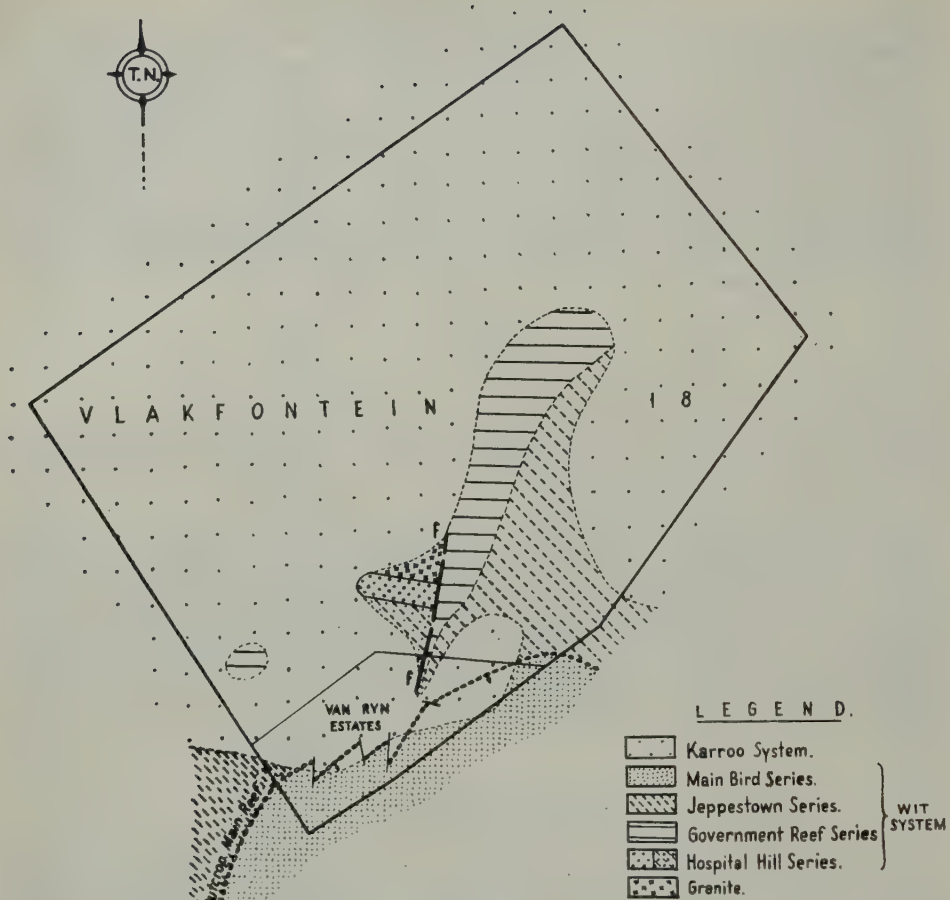


Fig. 2

In the southern portion of the Far East Rand also there are only a few isolated exposures of the Upper Witwatersrand System, although further to the south in the vicinity of Nigel, and to the east and south-east of Heidelberg, the full succession of both the Upper and Lower Witwatersrand Systems is exposed. The formations dip in directions varying from west to north to north-north-east, but the regional dip is to the north-west. In a few areas to the east and south-east of Nigel the base of the Witwatersrand System is exposed. In most of these exposures the Witwatersrand System rests on the Old Granite, although on the farm Rietfontein 72 it rests on a small remnant of the schists of the Primitive System. These exposures of the Witwatersrand System show distinct signs of folding, especially to the east and south of Nigel. On the farm Holgatfontein 127 there is an anticline, the axis of which trends to the north-west. In the area to the south-east of Nigel all the sediments of the Witwatersrand System have been folded and form one large anticline with minor folds superimposed on its crest. The crest of this anticline trends from 40° south of east to 40° west of north and pitches to the north-west at an angle of about 15°.

Further to the south-west on the farm Lagerspoort 310 there is a gentle anticline at the Main Reef Horizon which trends from south-east to north-west, and has a gentle pitch to the north-west. This anticline dies out in the upper portion of the Upper Witwatersrand near the town of Heidelberg and is not present near the contact between the Upper Witwatersrand and Ventersdorp Systems (see Fig. 10).

The basic amygdaloidal lavas of the Ventersdorp System overlie the Witwatersrand System on the Far East Rand and are present in the western and south-western portions of the area under discussion, but only isolated patches of these lavas are exposed on the surface, being covered by younger formations, although to the south of the area roughly south and west of Heidelberg these lavas are exposed over extensive areas.

The Black Reef and Dolomite Series of the Transvaal System overlie the Witwatersrand and Ventersdorp Systems unconformably.

Over most of the eastern portion of the Far East Rand the Transvaal System is disposed in the form of shallow basins (Mellor, 1921) but has a regional dip of about 5° to the east. See Plate III, sections III, IV, VI.

In the south-western and western portions of the Far East Rand the Transvaal System overlies the Ventersdorp System unconformably and in some places rests on the Witwatersrand System. Once again the Transvaal System is disposed in the form of shallow basins but there is a general, very gentle regional dip to the west.

To the north of the Far East Rand, on the farms Tweefontein 167 and Zesfontein 170, the Transvaal System rests on the Old Granite and has a gentle dip to the east.

The Dwyka and Ecca Series of the Karroo System overlie all the above systems unconformably and at least 60 per cent of the surface area of the Far East Rand is covered by these beds. It is only where these beds have been stripped that the underlying foundations are exposed. The beds of the Karroo System, for all practical purposes, lie horizontally.

The so-called Karroo dolerites, both sills and dykes, intrusive into the Karroo System, form ridges and tables on the surface, especially to the west of the centre of the Far East Rand in the vicinity of South African Land and Exploration Company, Ltd.

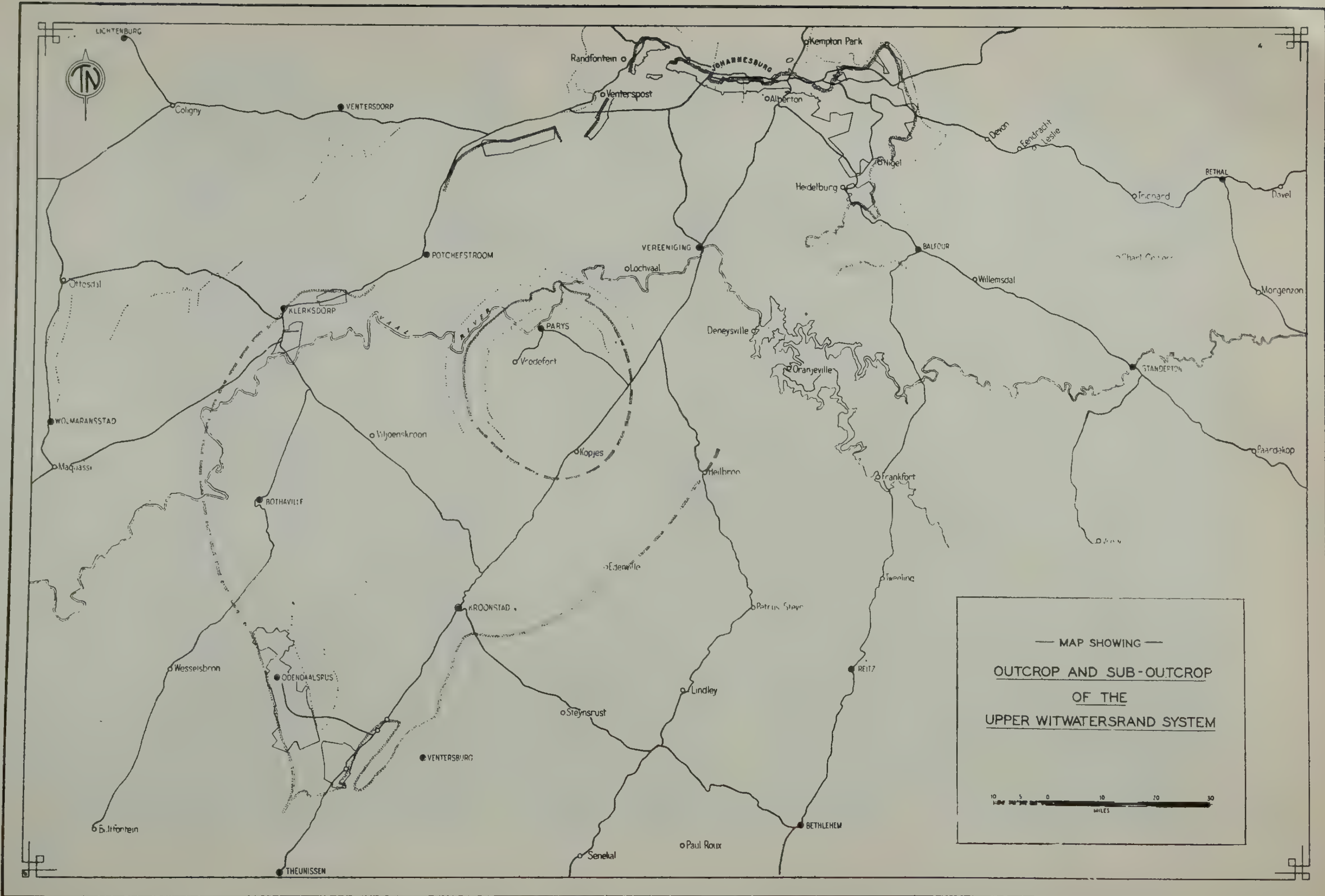
In the vicinity of all streams there is a thin covering of alluvial sand.

The entire country can be classed as grass-land, although when mining operations were started on the Far East Rand large numbers of gums and pines were planted and a good deal of timber has been obtained from these plantations. In many parts oaks, planes and poplars thrive and in fact most types of trees grow well on the Far East Rand.

Extent of Upper Witwatersrand System

Recent investigations in the Orange Free State and Western Transvaal, mainly by means of boreholes but also by geophysical methods, have shown that the Upper Witwatersrand System is more extensive than originally estimated.

The area over which the beds of the Upper Witwatersrand System are now known to be present is roughly egg-shaped, with the long axis trending from Springs district in the north-east, to Winburg district in the south-west, roughly a distance of 200 miles.



— MAP SHOWING —
OUTCROP AND SUB-OUTCROP
OF THE
UPPER WITWATERSRAND SYSTEM
10 5 0 10 20 30
MILES
LICHTENBURG
Coligny
VENTERSDORP
Randfontein
Venterspost
JOHANNESBURG
Kempton Park
Alberton
Devon
Eendracht
Leslie
Nigel
Heidelburg
BETHAL
Davel
BALFOUR
Willemsdal
Morgenzon
VEREENIGING
Lochvaal
POTCHEFSTROOM
Ottosdal
KLERKSDORP
VAAL RIVER
PARYS
Vredefort
Deneysville
Oranjeville
STANDERTON
WOLMARANSSTAD
Viljoenskroon
Kopjes
Heilbron
Frankfort
Paardekop
Maquassi
ROTHAVILLE
Edenville
Tweeling
KROONSTAD
Wesselsbron
ODENDAALSRUS
Lindley
REITZ
Steynsrust
VENTERSBURG
BETHLEHEM
Paul Roux
Senekal
THEUNISSEN

The short axis of the area is at its widest on a line between the Klerksdorp district in the north-west and the Heilbron district in the south-east, approximately a distance of 100 miles.

Along the periphery of this egg-shaped area the beds of the Witwatersrand System dip towards the centre, thus forming a large basin. The Parys-Vredefort Dome is situated in, and slightly north of, the centre of this egg-shaped basin.

Over the greater portion of the egg-shaped basin the Upper Witwatersrand System lies buried beneath younger formations.

The Pongola Series, which is situated about one hundred miles to the east of this Witwatersrand System basin, has not been correlated definitely with the Witwatersrand System, and although very likely of the same age it is certainly not continuous with it and does not fall within the scope of this discussion.

Regional Nomenclature of the Witwatersrand System

Various names have been given to different areas underlain by the Witwatersrand System, and refer for the most part to sectors of the egg-shaped basin where the Upper Witwatersrand System is exposed and its sub-outcrop is being mined.

The north-eastern sector of the basin is called the "Far East Rand" or "East Rand Basin", and the towns of Brakpan, Benoni, Springs, Dunnottar and Nigel are situated in this area. Along the northern edge of the basin in the vicinity of Boksburg the sector is called the "East Rand", and the sector extending roughly from 10 miles east to 10 miles west of Johannesburg is called the "Central Rand". To the west of these, in the vicinity of the towns of Luipaards Vlei, Krugersdorp and Randfontein, the sector is called the "West Rand". South-west of the West Rand is the West-Wits Line, an area extending from near Middelvlei Siding to Gats Rand Siding near the Mooi River. This sector is being called the "Far West Rand". In the western sector in the vicinity of Klerksdorp, are the "Klerksdorp Gold Fields", and in the south-western portion of the egg-shaped basin in the vicinity of Odendaalsrus, the sector is called the "Odendaalsrus" or "Orange Free State Gold Fields".

References to Previous Investigations relating to Structural Features of the Far East Rand.

The structure of the East Rand Basin has never been described in any detail. Hatch (1904) and Henderson (1905) came to the conclusion that the area was the eastern extremity of the Witwatersrand Syncline with the axis pitching west, Mellor (1915) mentions "the extensive and comparatively shallow synclinal basin of the Far East Rand", and Rogers (1921) "the wide syncline with east and west trend". Reinecke in 1927, spoke of the "main syncline" and the Nigel and Geduld anticlines which "are secondary folds on the flanks of the main syncline". Ellis (1943) in describing a tear fault states that this occurred at about the same time as the beds of the East Rand Basin were being folded. Fox (1939) in describing the area to the east of the East Rand Basin mentions the "south-east trending cross-folds superimposed on the south limb of the East Rand syncline".

Both the basin or syncline and the cross-folds have been recognised, and for some time it has been accepted that the East Rand Basin is the eastern extremity of a broad syncline, the northern limb of which extends over the whole length of the "Central Rand" and "East Rand" portions of the Witwatersrand System.

It has been assumed, generally, that the broad syncline was due to subsidence, although du Toit (1939) in a short description speaks of "the further tilting up of the Witwatersrand succession, to high angles, accompanied by thrusting on a considerable scale" during and following the Ventersdorp Eruptions.

Recently Brock (1950), in describing the attitude of the sediments of the Witwatersrand System, which presumably includes the Far East Rand, stated: "The inclination of the beds is the result of the relative upward movement of the granite hubs round the rim of the geosyncline or alternatively, the subsiding of the basin relatively to the rim; probably both"; and Truter in a discussion on this paper, states: "this basin is not a geosyncline but a synclinorium, and as far as I am aware it is now generally agreed that both kinds of structure result from compression and not from subsidence or uplift".

No attempt has been made to describe the structure of the Far East Rand in any detail or to explain the origin of the so-called "cross folds", and no map of the relief of the Main Reef has as yet been published.

II STRATIGRAPHY OF THE FAR EAST RAND

Witwatersrand System:

The sediments of the Witwatersrand System were deposited unconformably on a plane surface composed of the Old Granite and in a few places on remnants of the schists of the Primitive System.

The sediments of the Witwatersrand System, which have been divided into two main subdivisions, namely, the Upper and the Lower, have been described in great detail in numerous works, and the most complete and standard are those of Mellor (1911, 1915, 1917), Rogers (1921, 1922) and Nel (1927). The following, however, is a brief summary of these two main subdivisions of the Witwatersrand System in the East Rand Basin:—

(i) The Lower Witwatersrand varies in thickness from 12,000 feet to 9,000 feet, and consists of alternating zones of shales, sandy shales and quartzites with one thin interbedded lava flow near the top of the division, namely, the Jeppestown Amygdaloid. These sediments are very much less homogeneous and less competent than the overlying Upper Witwatersrand horizons.

(ii) The Upper Witwatersrand System varies in thickness from 5,500 feet to 3,000 feet and even less in the north-eastern portion of the Far East Rand, and consists of compact quartzites with some gritty and conglomerate bands with a zone of arenaceous shales (Kimberley Shales) from 400 feet to 200 feet thick situated from 2,000 feet to 1,000 feet above the base of the division, and a zone of basic lavas (Bird Amygdaloid) from 150 to 200 feet thick situated from 900 feet to 500 feet above the base of the division.

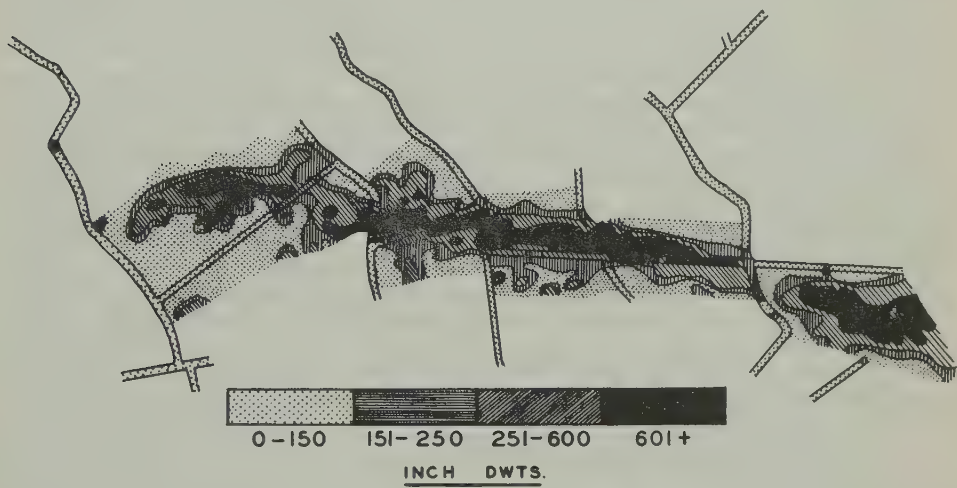
The Main Reef is situated at the base of the division, that is at the base of the Upper Witwatersrand. With the exception of the Kimberley Shales and the Bird Amygdaloid, the whole of the Upper Witwatersrand forms a very compact and competent sedimentary series. The Upper Witwatersrand rests with regional conformity on the Lower Witwatersrand, although there was a break in the deposition of sediments between these two divisions in portions of the Far East Rand. The Jeppestown Series, predominantly shales, (the uppermost portions of the Lower Witwatersrand), were in places eroded and slightly folded and the hollows formed both

by erosion and folding were filled with quartzites, grits and shales. These "filled hollows" are of economic importance and are known as "erosion channels" and "Footwall Reefs" (Carleton Jones, 1936). When this locally planed pre-Upper Witwatersrand surface was once again inundated and the hollows and erosion channels filled, the Main Reef conglomerates were deposited. This unconformable condition applies only to very isolated patches, and over most of the Far East Rand the Upper Witwatersrand follows quite conformably on the Lower Witwatersrand System.

The Main Reef is a bed of conglomerates which varies in thickness from a single layer of small pebbles to a band over five feet in thickness, but is usually about three feet thick with pebbles usually one inch in diameter. The Main Reef is the principal conglomerate mined in the Far East Rand and about 90 per cent of the gold of this area has been obtained from this horizon.

The composition of this compact conglomerate bed has been described in numerous publications, and the works of Young (1917), Pirow (1920), Mellor (1913, 1915), Reinecke (1927, 1930) and Nel (1927) cover most of the significant characteristics, of this "Banket".

One important aspect of the Main Reef which has a bearing on the tectonic history of the Far East Rand, namely, the "Reef Shoots", must be mentioned. In all mining operations on the Far East Rand it has been found that the gold content in the Main Reef varies considerably and that the zones of both high and low run in definite lines or "shoots".



PORTION OF A FAR EAST RAND MINE
SHOWING MAIN REEF PAY SHOOT
SCALE 1:2000

Fig. 3

The width of these shoots is often highly variable, occasionally bellying out to hundreds of feet and sometimes narrowing down to ten or less feet. The thickness of the conglomerate also varies but the variation is slight and is measured in inches.

The length of the shoots is also variable, can often be traced for distances of ten miles and more, but usually extends from two to five miles. These shoots generally trend from north-west to south-east, but gradually fan out, especially along the northern portion of the Far East Rand, and in the north-eastern portion of the Far East Rand trend from west to east.

A very important aspect of mining in these parts is to trace these "shoots", and most mines now prepare what are called "Colour Contour Maps" of the gold values obtained in assays of the reef in all stoping and development. Thus for instance on some mines all gold values varying from zero to 100 inch-dwts. are coloured green, values varying from 100 inch-dwts. to 150 inch-dwts. yellow, and so on. The "Value Contour Maps" have proved beyond doubt the existence of definite trends or "shoots" in the gold values, and these "shoots" sometimes run parallel to the trend of the "Cross-folds" but more often trend across the folds and transgress from anticline to syncline. An example of a "payshoot" taken from the plans of one of the mines of the Far East Rand is illustrated by Fig. 3.

What is important as far as this discussion is concerned is the fact that it is possible to determine the horizontal displacement of these shoots when they are cut by certain faults which trend roughly east-west.

Ventersdorp System:

In the Far East Rand the Witwatersrand System is overlain conformably by basic lavas of the Ventersdorp System. This conformity applies to the Central and Far East Rand sectors of the Witwatersrand, but on the West Rand the lavas are unconformable to the Witwatersrand System (Pelletier, 1937). The apparent conformity of these two systems on the Far East Rand is remarkable, as there obviously elapsed a considerable period of time between the deposition of the sediments of the Witwatersrand System and the pouring out of the lavas of the Ventersdorp System on the Far West Rand and other areas, but on the Far East Rand there is very little evidence of erosion and the contact between the two systems is clear-cut. In certain parts, especially in the vicinity of Heidelberg, a tuff is present near the contact, and the following is a description of these deposits by Rogers (1922).

> "The tuff is persistent throughout the area mapped and the frequent presence of quartz grains and the occasional layers of small pebbles in it, prove that there was an intermingling of ordinary sediments with the tuff during deposition. The arenaceous deposits were overwhelmed by the volcanic, but there was no sudden and complete cessation of the supply of sand, and the existence of the passage beds over a wide area is very good evidence of conformable succession of the Ventersdorp rocks to the Elsburg in the Heidelberg district".

In boreholes and shafts put down in the vicinity of the Sub Nigel Ltd., West Vlakfontein G.M. Co. Ltd., and Vlakfontein G.M. Co. Ltd., it is possible to distinguish individual lava flows, and the strikes and dips of these flows are the same as those of the underlying Elsburg beds of the Witwatersrand System.

The lavas occur only in the south-western portion of the Far East Rand, and over the remaining and major portions of this area both the lavas and varying portions of the Witwatersrand sediments were eroded away prior to the deposition of the sediments of the Transvaal System.

Transvaal System:

On the Far East Rand the Black Reef Series and Dolomite Series of the Transvaal System rest unconformably on a floor composed of either lavas of the Ventersdorp System or sediments of the Witwatersrand System, and for the most part are horizontal or dip at very flat angles. These deposits have been described in detail by Rogers (1922), Mellor (1915) and Nel (1933). The Black Reef Series is a thin zone seldom more than 40 feet thick but occasionally up to 200 feet thick, composed of darkish quartzites with some shaly bands, and the Dolomite Series consists of dolomite with bands and lenses of chert.

The Karroo System:

On the Far East Rand, the Dwyka and Ecca Series of the Karroo System rest unconformably on a floor composed of either Witwatersrand, Ventersdorp or Transvaal Systems, and on the Old Granite, and fill hollows that were present in the Post-Transvaal System landscape. These deposits have been described in detail by Rogers (1922), Mellor (1915) and Nel (1927). The Dwyka Series is usually represented by a thin band of tillite seldom more than four or five feet thick and the Ecca Series by sandstones with shale bands, and in places seams of coal.

Igneous Intrusions in the Far East Rand Basin.

In the Far East Rand there are a number of related dykes and also sills, and the following is a brief discussion on these intrusives:

Dykes striking north-west and south-east.

Ellis (1946) has given a description of three of the largest dykes which he shows are of Karroo Age and which strike roughly north-west and south-east, and dip almost vertically. The most easterly of these dykes runs from Modderfontein East, Ltd., through East Daggafontein Mines, Ltd., and extends into Marievale Consolidated Mines, Ltd. The second dyke is parallel to the first and extends from Van Ryn Gold Mines Estate, Ltd., to Springs Mines, Ltd. and is most likely present in The Nigel Gold Mining Company, Ltd. The third runs through Van Dyk Consolidated Mines, Ltd., but its southern extension has not been exposed by mine workings (see Plate II).

These three dykes cut through all other dykes and faults and all formations from the Karroo to the Witwatersrand System.

Ilmenite-Diabase Dykes.

These dykes follow fissures and faults both in the horizontal and vertical planes but their general trend is north and south. Ellis (1946) considers these dykes to be of late Transvaal System age.

Acid Intrusions.

Acid intrusions are very rare but a few of post-Transvaal System age are known.

On Marievale Consolidated Mines, Ltd., Vogelstruisbult Gold Mining Areas, Ltd., The Sub Nigel Ltd., and Vlakfontein Gold Mining Company Ltd., acid rocks are encountered at the Main Reef horizon and trend east and west occupying shear zones.

Other Dykes.

Numerous smaller dykes are found cutting the Main Reef horizon but very

few persist for any great distance. These dykes are of various ages, the majority probably of late Ventersdorp Age, and they strike in all directions, but usually follow fault planes or fissures.

Sills.

A sill of major importance in mining operations on the East Rand is one called the "Old Lady" (1940). This sill, a quartz-dolerite, normally occurs in the Jeppestown Shales but often breaks through the Main Reef at steep angles and then follows the bedding in the quartzites some hundreds of feet in the hanging of the Main Reef. Where the "Old Lady" breaks through the Main Reef the reef is displaced downwards from 200 to 180 feet, which is the thickness of the sill. The largest known area where the sill has domed over the Main Reef is one that covers a large portion of Marievale Consolidated Mines, Ltd., the eastern portion of Vogelstruisbult Gold Mining Areas, Ltd., the southern portion of East Daggafontein Mines, Ltd., the eastern portion of Springs Mines, Ltd., and the central portion of Daggafontein Mines, Ltd.

A further area where the sill has domed over the Main Reef is found in the northern portion of Springs Mines, Ltd., and the southern portion of New State Areas, Ltd.

The "Old Lady" is known to cut through the Main Reef in other areas but insufficient development has been carried out to trace the intersection of sill and Main Reef over any distance.

Other basic sills are known to be present in the Kimberley-Elsburg Series especially in the vicinity of the Kimberley Shales.

Mapping by Rogers in the Heidelberg district has shown that many quartz-dolerite sills are present in the Lower Witwatersrand System.

Faults.

Hundreds of faults are found in the Far East Rand Basin. They traverse the Main Reef horizon and strike in all directions, but in a number of cases there is a remarkable amount of parallelism. One set of faults strikes roughly east and west, another roughly north and south, and these two sets of faults are responsible for the major displacements both in the horizontal and vertical planes (see Plate II). The faults of these two sets dip at steep angles, usually between 75° and vertical. There are two additional sets of faults which strike roughly from north-east to south-west and from north-west to south-east, but these are rather impersistent and usually end on the north-south and east-west faults. All these faults will subsequently be described in more detail.

III SURVEY PRACTICE AND MAPPING ON THE WITWATERSRAND MINES

All the mines on the Witwatersrand are surveyed in great detail both on surface and underground. The Union Government has published a number of surface maps of the Witwatersrand and the most complete of these is one done on a scale of 1 : 50,000 and this shows surface contours at 50 foot intervals. Most mines have surface contour maps on a scale of 1 : 2,000 or 1 : 5,000 giving greater detail than the Government maps and, depending on the mine, surface contours at 1 foot or 5 foot intervals.

All underground workings are surveyed in great detail and innumerable points are co-ordinated on all haulages, drives, raises, winzes, vertical shafts, incline shafts, etc., and plans and sections are made of these underground workings to a scale of 1 : 100 or 1 : 200.

All mines, by law, have very large plans made of the whole mine on a scale of 1 : 1,000 and these plans are kept up to date. In addition to the above plans, most mines prepare others to the scale of 1 : 2,000, 1 : 5,000 and 1 : 10,000. On plans with scales up to 1 : 1,000 the number of each surveyed peg is plotted and in many instances the elevations of these pegs are plotted on the plans as well. The co-ordinates and elevations of all pegs are logged in books, and it is a simple matter to find all particulars that are not plotted on the plans.

The co-ordinates of all pegs are calculated from standard base lines and do not need any explanation.

The elevations of all surface points are also quite normal and are calculated on the usual sea level datum.

The elevations of all mine workings on the Witwatersrand are different from the usual, however, as datum or zero is 6,000 feet above sea level. All elevations on and in mine workings are calculated relative to this base.

The elevations of all mine workings are below this datum line and are hence negative (—), which distinguishes them from surface contours and elevations.

The following examples will explain the above:—

The collar or surface elevation of No. 1 shaft on Vlakfontein Gold Mining Company, Ltd., is —629·44 feet, i.e. 629·44 feet below datum. As this datum is 6,000 feet above sea level the collar of this shaft is 6,000 ft. — 629·44 ft. that is 5,370·56 feet above sea level.

Again the reef elevation in the same shaft is —7,276·44 ft. that is 7,276·44 feet below datum. As this datum is 6,000 feet above sea level, the reef is 1,276·44 feet below sea level.

The Main Reef Contour Plan.

The writer, during the last fifteen years, has on numerous occasions visited each of the mines of the Far East Rand and during this time compiled contour plans of the Main Reef of each of these mines. In some notable cases contour plans have been prepared by resident mine geologists during recent years, and copies of these plans were kindly supplied by these geologists with the permission of the respective Mine Head Offices.

In compiling these reef contours the following procedure was adopted: Plans of the mine workings were obtained from each mine, usually on a scale of 1 : 2,000 or 1 : 2,500 and in some instances on a scale of 1 : 5,000. These plans show in detail all the underground workings, indicate all work that has been done on "on reef" and on all "off reef" development, and in many instances indicate the presence of dykes, sills, faults, etc. On these plans innumerable reef elevations were then plotted, especially when development had been done by means of haulages which undulate, and on winzes and raises. When development had been done by means of contour drives naturally it was not necessary to plot so many elevations, and elevations at about 300 foot intervals were found to be sufficient. Known faults and dykes were plotted next and then reef contours were drawn usually at 50 foot intervals, but occasionally, when some detail in structure was obscure, at 10 foot intervals. Invariably it was necessary to visit each mine a number of times to obtain further details, and often it was necessary to go underground to verify data before the contour plan could be completed. This applied particularly in the correlation of some dykes and faults.

When the reef contours of a mine were completed, and very often this was done on four or five different plans, these contours were transferred to a plan of a scale of 1 : 10,000. The 1 : 10,000 plans were then photostated down to a scale of 1 : 50,000 and tracings made of these photostats on to the composite map of 1 : 50,000 of the whole of the East Rand Basin. It is impossible to visualise the structure of the Main Reef in the East Rand Basin on large mine plans, as the broad outlines are obscured completely by minor folds, faults and displacements, but on the scale of 1 : 50,000 the structure of the area as a whole can be seen clearly (see Plate II).

It must be remembered that all elevations are below the datum for the Witwatersrand mines, so that the contour marked 1,000 feet is near the surface and outcrop or sub-outcrop, and roughly between 300 and 400 feet below surface, the contour marked 5,000 is deeper and roughly 4,300 feet or 4,400 feet below surface, and the contour marked 8,000 feet is still deeper and roughly 7,300 feet or 7,400 feet below surface, depending on the surface elevation.

The outcrop and sub-outcrop have been marked with a heavy dotted line, and although there are no mine workings along the eastern fringe of the basin, the position of the sub-outcrop has been obtained fairly accurately from the results of numerous boreholes. The reason for the lack of detail along the eastern fringe is that boreholes have shown that the gold content of the reef in this part of the basin is rather low and sporadic, and very little or no mining has been carried out.

The blank spaces on the western side of the map are where the reef is for the most part deep and for this reason has not to date been developed or mined, although here too a number of boreholes have been drilled and some of these have shown that the reef has a satisfactory gold content.

IV MAIN STRUCTURAL FEATURES OF THE COUNTRY ADJOINING THE FAR EAST RAND

The Far East Rand forms the north eastern corner of the large egg-shaped basin of the known Witwatersrand System.

To the north-west of the Far East Rand is the large Old Granite mass which extends roughly from Johannesburg to Pretoria, and it is on a portion of this Old Granite that the Witwatersrand beds were deposited.

To the east of the Far East Rand the Witwatersrand System is covered by both the Transvaal and Karroo Systems, and various mining ventures have explored this eastern sector in the hope of finding further extension of the Main Reef, as for instance "B" Far East Rand Ltd., No sediments of the Upper Witwatersrand have been found to the east of the Far East Rand Basin up to the present time, but drilling has shown that sediments of the Lower Witwatersrand are present, apparently much disturbed by folding and faulting.

To the south of the Far East Rand Basin another granite mass occurs. This granite is once again the Old Granite on which the Witwatersrand System was deposited, and from the few exposures appears to be dome shaped.

Structure of Far East Rand

In the following description it is the Main Reef Horizon of the Witwatersrand System that is referred to. Although the remaining beds of the Upper Witwatersrand

System are roughly conformable, only one other sedimentary layer, the May Reef, one of the so-called Kimberley Reefs, which lies between 1,000 feet and 2,000 feet above the Main Reef, has been mined to any extent.

Sufficient work has been carried out on the May Reef to show that apart from minor irregularies it has the same regional structural features as the Main Reef.

Detailed mapping of the Main Reef Horizon, and to a lesser degree the May Reef Horizon, and the surface geology, show two principal structural features, namely:—

(*a*) A basin.

(*b*) Major folds.

The following are descriptions of these two features.

(*a*) *The Far East Rand Basin.*

The outcrop and sub-outcrop of the Main Reef on the Far East Rand is very irregular, but may be considered as egg-shaped with the ends of the long axis distorted. The long axis measures about 19 miles and trends approximately 20° west of north. The short axis is about 15 miles long (see Fig. 4 (*a*).

The Main Reef in general dips south-south-east from the northern flank, west-south-west from the eastern flank, and north-north-west from the southern flank of the basin. The western side of the basin presents, as far as is known, no easterly dip, and west of the western portion of Van Dyk Consolidated Mines, Ltd., and of the farm Withok 149, dips fairly steeply to the west and the Central Rand Basin.

At its deepest, in the vicinity of West Springs, Ltd., Vlakfontein Gold Mining Company, Ltd., and Withok joint boundary, the Main Reef is about 8,000 feet deep (i.e. below surface).

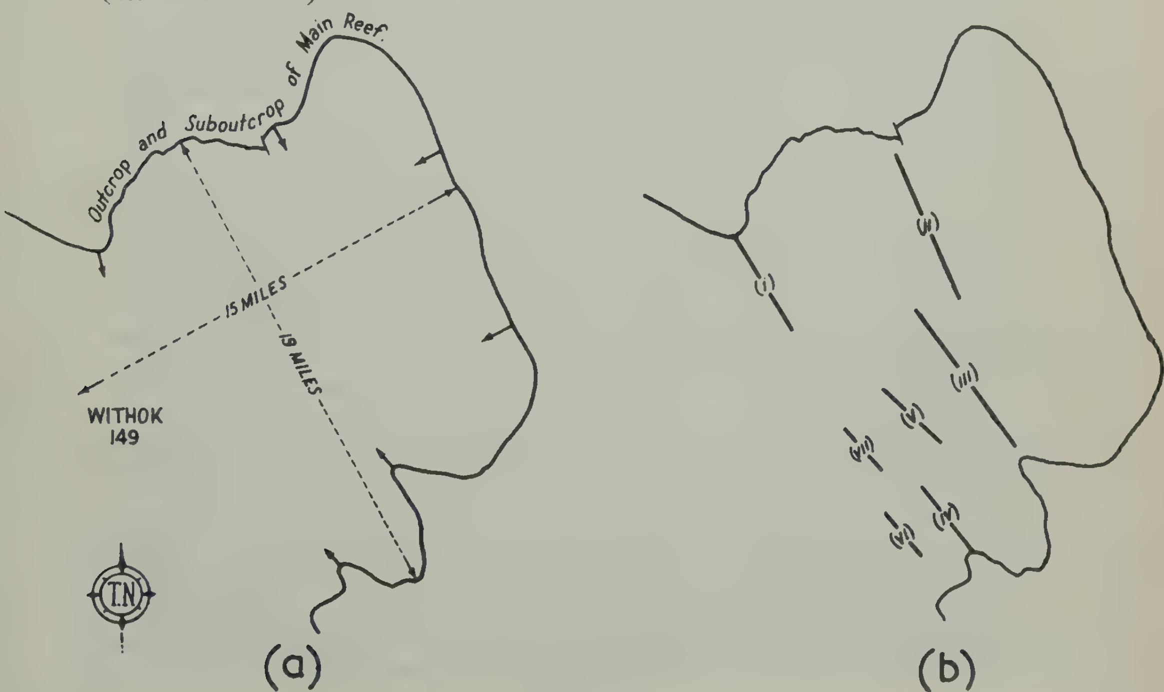


Fig. 4

(*b*) *Major Folds.*

The mining of the Main Reef has shown that there are four major anticlines in the Far East Rand Basin and these anticlines are also present in all the Upper Witwatersrand Beds.

The axes of all the folds have a north-north-west, south-south-east trend and are arranged en echelon and roughly parallel to the eastern rim of the basin (see Fig. 4 (*b*)).

The following is a list and description of the anticlines, which are named after the mines on which they occur:—

(i) Van Dyk-Brakpan anticline.
(ii) Modder "B"-Geduld anticline.
(iii) Marievale-Vogelstruisbult-Springs anticline.
(iv) Nigel-Sub Nigel anticline.
(v) Sub Nigel-Vlakfontein anticline.
(vi) Spaarwater anticline.
(vii) West Vlakfontein-Vlakfontein anticline.

Anticlines (iv), (v), (vi) and (vii) really form one large anticline, which may be called the Nigel-Sub Nigel Spaarwater-West Vlakfontein-Vlakfontein composite anticline.

(i) *The Van Dyk-Brakpan Anticline.*

This anticline is situated on Van Dyk Consolidated Mines, Ltd. and the western portion of Brakpan Mines, Ltd. On the eastern flank are sited the mines of New Kleinfontein Co., Ltd., and Van Ryn Deep, Limited. Most of this anticline has been exposed by underground development and stoping, although the lower levels of the south-western flank have not as yet been developed completely.

The crest of the anticline trends 40° west of north. The crest or crest line has a variable pitch to the south-south-east. From the sub-outcrop to a horizontal distance of about 3,000 feet down dip, the pitch averages about 16°. From this point for the following 8,000 feet the pitch averages about 5° and beyond this point the pitch gradually steepens to 25° and more. The horizontal length of the anticline at the Main Reef horizon is roughly 22,000 feet or just over 4 miles.

The south-western flank of the anticline is very regular along its strike. This flank dips to the south-west and the dips are gentle from the crest to a horizontal distance of 2,000 feet from the crest, but from this point steepen sharply, to angles of 60° and more, to the south-west.

The eastern flank is well defined but not regular, as a number of minor folds occur along it. The crests of these minor folds are roughly parallel to the crest of the main anticline. Owing to this minor folding the dips of the eastern flank vary considerably from gentle angles near the crest to angles varying from 15° to 40°, but as a whole the flank can be said to dip at approximately 20° to the east.

In the southern portion of this flank the dips gradually flatten and in the south-eastern corner of Brakpan Mines, Ltd. and on South African Land and Exploration Company, Ltd. this flank disappears and the Main Reef has a strike practically due east and west.

The axial plane of this anticline strikes 40° west of north and dips to the east-north-east at approximately 70°. This axial plane could not be calculated more accurately as only one thin layer of the anticline, namely, the Main Reef, was mapped.

The Van Dyk-Brakpan anticline may be classed as a variable pitching asymmetrical anticline with the axial plane dipping at approximately 70° to the north-east. The crest pitches to the south-south-east at angles varying from 5° to over 25°.

This is therefore a 20° overfold from east to west as deduced from the dip of the axial plane.

This anticline occupies the north-western extremity of the Far East Rand, and to the west of this anticline the formations apparently dip steadily into the large Central Rand Basin and all signs of folding disappear as far as is shown by present underground work.

The outcrop of the whole of the Upper Witwatersrand from Van Dyk Consolidated Mines, Limited to Durban Roodepoort Deep, Limited shows no signs of north-west to south-east trending folds. The sediments over this length of 20 miles dip to the south, and strike east-west with the strike very slightly convex to the south.

(ii) *The Modder "B"-Geduld Anticline.*

This anticline is situated on Modderfontein "B" Gold Mines, Ltd., Geduld Prop. Mines, Ltd., Modderfontein East, Ltd., and portions of East Geduld Mines, Ltd., New State Areas, Ltd., and Springs Mines, Ltd.

The crest of the anticline at the Main Reef horizon starts at the outcrop of Modderfontein "B" Gold Mines, Ltd., and trends 30° west of north.

The anticline has a fairly regular pitch to the south-south-east at angles varying from 5° to 10°.

The eastern flank is regular along its strike and dips to the east. The dips near the crest are gentle and gradually increase down the dip to a maximum of about 23°, and then again flatten out towards the trough of the syncline. From the crest of the anticline to the trough of the syncline on the east, the horizontal distance is approximately 10,000 feet and the difference in elevation approximately 1,500 feet.

The south-western flank also is regular along its strike, and dips to the south-west. The dips near the crest are gentle, gradually increase down-dip to a maximum of about 24° and then decrease to low angles further to the south-west.

This flank does not end in a clearly defined trough as does the eastern flank, but rather in a large open or wide, slightly undulating, plane. This condition applies to the lower or south-eastern three-quarters of the flank.

The distance along the crest from the outcrop to the 4,500 foot level where the anticline fades out into the Daggafontein syncline is about 41,000 feet, or slightly less than 8 miles.

This is a remarkably regular anticline and may be classed as symmetrical with the axial plane vertical and the crest or axial line pitching comparatively regularly to the south south-east at angles varying from 5° to 10°.

The anticline is crossed by numerous faults; one of these, a thrust fault which runs through the lower levels of East Geduld Mines, Ltd. in an east to west direction, displaces the crest to the west on the north side of the fault, but the faults, including the above thrust fault, do not distort the regularity of the anticline to any great degree.

(iii) *The Marievale-Vogelstruisbult-Springs Anticline.*

This anticline starts from the sub-outcrop on Marievale Consolidated Mines, Ltd., continues through roughly the centre of Vogelstruisbult G.M. Areas, Ltd., through Springs Mines, Ltd., and dies out in a large flat basin near the northern boundary of the last-mentioned mine. The crest of the anticline trends 37° west of north.

The pitch of the anticline is to the north-west and varies from about 30° at the sub-outcrop, gradually flattening out to a distance of about 14,000 feet from the sub-outcrop. From this point to where the anticline fades out on Springs Mines, Ltd.,

the pitch varies from 3° to 4° to the north-west. The length of the anticline at the Main Reef horizon is roughly 50,000 feet or 9 miles, measured horizontally.

The crest of the anticline has been displaced by a large transverse tear fault which runs almost east and west, and on the north side of the fault the crest has been relatively displaced some 2,000 feet to the west. There are other faults parallel to the above fault but the displacements are negligible. In addition to these east-west faults there are minor faults some of which strike parallel to the crest of the anticline, others at right angles to the crest line and again others at a variety of angles to the crest line.

The north-eastern flank of the anticline has been exposed to a large extent by underground development. The dips down this flank vary quite considerably. Near the crest the dips are gentle, varying from 3° to 5°, except near the sub-outcrop where dips of 25° are common. Down-dip from the crest the dips gradually increase and reach their maximum about 2,500 feet from the crest. Here again the greatest dips of up to 30° are near the sub-outcrop, gradually decreasing till on Vogelstruisbult G.M. Area, Ltd., they are in the vicinity of 20° and further north are 10°, and less still in the northern portion of Springs Mines, Ltd.

The south-western flank has been exposed by underground development in the northern portions of the flank but the southern portion has not been developed to any great extent, although sufficient work has been done to give the general picture of this portion of the flank. Along almost the entire length of this flank there is a zone where the dips are from 60° to 70° to the south-west. Along the most southerly portion of this flank, that is on Marievale Consolidated Mines, Ltd., to the immediate east of Nigel G.M. Co., Ltd., this steeply dipping zone extends from the sub-outcrop to about — 2,500 foot level. Further north, about halfway across Vogelstruisbult G.M. Areas, Ltd., the steep zone lies between the elevations of — 4,500 feet and — 6,500 feet. In the south-eastern portion of Springs Mines, Ltd. this zone lies between the — 5,000 foot and — 7,000 foot levels, although in this area the dips are not quite so steep, varying from 50° to 60°. Further north, this zone gradually becomes less steep and in the north-western portion of Springs Mines, Ltd. the dips have flattened to 20° and less.

From the crest of the anticline to a horizontal distance of about 2,000 feet to the south-west, the dips are gentle and vary from 3° to 10° and it is beyond this position that the sudden steepening begins. This condition applies over most of the length of this flank, except for that portion near the sub-outcrop on Marievale Consolidated Mines, Ltd., where the dips are fairly steep from the very crest.

The axial plane of this anticline dips to the north-east at an angle of approximately 70°.

The Marievale-Vogelstruisbult-Springs anticline may be classed as an asymmetrical fold (Billings) with the axial plane dipping to the north-east.

The crest of the anticline pitches or plunges to the north-west and the crest line is gently concave.

The Nigel-Sub Nigel Spaarwater-Vlakfontein and West Vlakfontein Anticlines.

In the area occupied by the mines of Nigel G.M. Co., Ltd., The Sub Nigel Limited, Spaarwater G.M. Co., Ltd., Vlakfontein G.M. Co., Ltd., and West Vlakfontein G.M. Co., Ltd., there are a number of anticlines, many of them over three miles long and situated en echelon to each other, but which together form one major anticline, the north-eastern flank of which terminates in a well-defined syncline whose trough trends south-east to north-west and pitches to the north-west.

The western flank of this major composite anticline is, as far as is known, the western edge of the East Rand Basin where the sediments dip down into the main

Central Rand syncline. The following are descriptions of the individual anticlines forming this composite anticline which is known to extend over a length of at least 8 miles and is probably over 6 miles wide at the Main Reef horizon.

(iv) *The Nigel-Sub Nigel Anticline.*

This anticline on the Main Reef horizon starts from the outcrop on Nigel G.M. Co., Ltd., continues into The Sub Nigel, Ltd., and passes over into a syncline in the lower levels or northern portion of this mine.

Most of the anticline has been exposed by underground development and only a small portion of the lower levels of the north-eastern flank has not been developed completely.

The crest line of the anticline runs from south-east to north-west and is for all practical purposes a straight line on the horizontal plane. The length of the crest is approximately four and three-quarter miles.

The pitch of the anticline is to the north-west and is fairly regular averaging about 12°.

The western flank dips to the west at angles seldom exceeding 20° and usually in the vicinity of 15°. The dip gradually lessens in a northerly direction and this flank loses its identity, disappearing into a syncline in the northern portion of The Sub Nigel, Limited. The horizontal distance from crest to trough along the flank is roughly one mile.

The north-eastern flank dips to the north-east at angles varying from about 10° near the crest to about 20° down the major portion of the flank. The dips are on the whole steeper near the outcrop on Nigel G.M. Co., Ltd., and in places are up to 30°, and also near the central portion of The Sub Nigel, Limited there is a small section where the dips on this flank are steep, with dips in places up to 45°. The horizontal width of this flank from crest to trough measured at right angles to the crest line is 10,000 feet or nearly two miles. The anticline is not a very regular one as both along the crest and on the flanks there are minor folds and puckers.

The anticline is cut by a number of east-west transverse tear-faults, also by a large north-east to south-west trending displacement which is caused either by a normal fault or a displacement of the Main Reef zone by a large 300 foot thick dolerite sill, which transgresses from the hanging wall to the footwall of the reef along this line. It has not been proved which of these caused the displacement. The Main Reef on the south-east side of this displacement has dropped roughly 300 feet relative to the Main Reef on the north-west side.

There is one other large displacement in this anticline, namely, one that runs roughly from north to south for a distance of nearly two miles, then turns to the south-west for about one mile and finally swings off in a westerly direction. This displacement, which varies from 350 feet to 500 feet, is known to be due to a sill, transgressing from the hanging wall to the footwall of the Main Reef, but not affecting the Kimberley Reefs in the area near the common boundary of The Sub Nigel, Limited and Nigel G.M. Co., Ltd. mines and also in the southern portion of Spaarwater G.M. Co., Ltd. That portion of the displacement that strikes roughly north and south, however, appears to be normal faulting with "loss of ground" but this has not been proved.

The anticline is crossed by a number of other faults which run in all directions, but for the most part with minor displacements.

The Nigel-Sub Nigel anticline is a gentle flexure or curve, and as only one bed, that is the Main Reef, in this curve can be examined it is not possible to locate the

axial plane with any degree of accuracy. The north-eastern flank is almost twice the length of the west flank but the dips of each of these flanks are very similar.

If the crest coincides with the axis, then the axial plane is probably vertical and the anticline may be called symmetrical. According to Stoces and White one of the requisites of a symmetrical fold is that the flanks are of equal length, but according to Sherbon Hills (1943) this is not necessary and the writer has based his nomenclature on Sherbon Hills' definitions.

The Nigel-Sub Nigel anticline may be classed as a symmetrical fold with a comparatively regular pitch to the north-west.

(v) *The Sub Nigel-VlakfonteinAnticline.*

This anticline is situated in the north-eastern portion of The Sub Nigel, Limited and the eastern portion of Vlakfontein G.M. Co., Ltd., and has not been exposed completely by underground development.

The crest of the anticline trends 34° west of north, but the extremities of this crest are not known exactly, although the south-eastern end must fade into the synclinal basin situated to the east of the centre of The Sub Nigel, Limited. The area where this end of the anticline fades is cut up by large faults with vertical displacements of between 300 and 500 feet, and only a small amount of underground development has been carried out on the Main Reef.

The position of the north-western end of the crest also is not precisely known, but it is present near the northern boundary of Vlakfontein G.M. Co., Ltd. The known length of this crest is approximately 22,000 feet or just over four miles.

The pitch of the anticline is to the north-west at angles varying from 5° to 8°.

The north-eastern flank dips to the north-north-east at angles varying from 10° to 25° but generally in the vicinity of 10°. The trough in which this flank ends has been disclosed in only one position near the common mine boundary between The Sub Nigel, Limited and Vogelstruisbult G.M. Areas, Limited. The horizontal distance from crest to trough in this position is approximately one mile.

The western flank dips to the west and north-west at angles varying from 5° to 10°. The horizontal distance from crest to trough is approximately one mile. This flank is very puckered and over almost its whole length has minor folds with crest lines running roughly parallel to the crest line of the main anticline. These minor folds are up to 4,000 feet long and with amplitudes averaging about 800 feet from crest to trough, and the elevation between these axes averages about 130 feet. This western flank is also crossed by a number of east-west trending tear-faults of minor but definite horizontal displacements.

The Sub Nigel-Vlakfontein anticline may be classed as a symmetrical fold with a pitch of between 5° and 8° to the north-west. The crest of the anticline trends 34° west of north. One rather unusual feature of the anticline is that the north-north-eastern flank is comparatively regular but that the western flank is much puckered by minor folds.

(vi) *The Spaarwater Anticline.*

This anticline is situated in the southern portion of Spaarwater G.M. Co., Ltd., and also in the area slightly south of the mine where the Main Reef occurs in sub-outcrop.

This anticline has been exposed only partly by underground development, but sufficient work has been done to make clear its general behaviour.

The crest runs from south-east to north-west and extends from the sub-outcrop south of Spaarwater G.M. Co., Ltd., through about the centre of Spaarwater G.M. Co., Ltd. and probably dies out in the northern portion of this mine, where it merges with the west flank of West Vlakfontein-Vlakfontein anticline.

The pitch of this anticline is to the north-west and from the limited amount of exposure underground appears to be fairly regular, averaging about 10°.

The northern flank dips to the north at angles of between 10° and 12° and is not very regular as it has been made uneven by a number of minor puckers. Also the Main Reef Horizon has been displaced by large dolerite sills which transgress from the hanging wall to the footwall of the reef. In addition, there are a number of minor faults.

The western flank dips in a westerly direction and has been exposed over a very limited area in which the Main Reef dips at angles varying from 5° to 10° near the crest, to 30° at a horizontal distance of 3,000 feet from the crest.

The western flank being steeper than the northern flank, at least in the portions exposed, indicates that the anticline is asymmetrical.

The Spaarwater anticline, on the information available, is probably asymmetrical with the axial plane inclined to the north-east. The crest of the anticline pitches to the north-west at low angles and trends from south-east to north-west.

(vii) *The West Vlakfontein-Vlakfontein Anticline*.

This anticline is situated in the eastern portion of West Vlakfontein G.M. Co., Ltd., the western portion of Vlakfontein G.M. Co., Ltd. and probably the northern portion of Spaarwater G.M. Co., Ltd., and although it has not been exposed to any great extent by underground development, sufficient work has been done to prove its existence and to give a general picture of its behaviour.

The crest of the anticline has been located at only one point, but sufficient work has been done to show that the crest line trends to the north-west and also that the pitch of the crest is to the north-west.

The north-eastern flank has been exposed over a length parallel to the crest of the anticline of approximately 20,000 feet or three and three-quarter miles, and further development will show that this flank is considerably longer, probably in the vicinity of 30,000 feet. The dips down this flank are to the north-east and are fairly regular varying from 15° to 20° except for the portion near the crest where the dips vary between 5° and 10°.

The south-eastern portion of this flank fades and merges with the Nigel-Sub Nigel anticline in the north-western portion of The Sub Nigel, Ltd. The north-western extremity of this flank is not known, but probably dies out in the western portion of the farm Withok 7.

The west flank has been exposed by underground development over a very limited area on West Vlakfontein G.M. Co., Ltd., but sufficient work has been done to show that this flank dips to the west at angles between 10° and 12°.

From the limited amount of work done along the west flank it is not possible to state with any degree of accuracy the details of the West Vlakfontein-Vlakfontein anticline, but the following deductions may be made:—

The anticline is at least 20,000 feet long, the crest extending from the northern portion of Spaarwater G.M. Co., Ltd., through West Vlakfontein G.M. Co., Ltd., and probably dying out in the north-eastern portion of Withok.

The north-eastern flank has a slightly steeper dip than the western flank, so that the axial plane probably has a dip to the north-east, with regard to the floor-slope of the basin.

The pitch of the anticline is to the north-west and, although this is not known accurately, is probably between 5° and 10°.

V THE BASINING OF THE WITWATERSRAND AND VENTERSDORP SYSTEMS ON THE FAR EAST RAND IN PRE-TRANSVAAL SYSTEM TIME.

In arriving at a possible cause of the deformation of the Witwatersrand System on the Far East Rand to its present basin-like shape, it is necessary to consider this area as being the north-eastern extremity of the large synclinal basin of the Witwatersrand System as a whole. Also, it is necessary to determine the age of the deformation and as far as possible the relative movements of the Witwatersrand System and the subjacent Old Granite and Primitive System.

The deformation of the Witwatersrand System is discussed under the following headings:—

- (*a*) The Johannesburg-Pretoria Old Granite as portion of floor of deposition for Witwatersrand and Ventersdorp Systems.
- (*b*) Old Granite to the immediate east and south-east of Nigel.
- (*c*) Pre-Witwatersrand System floor.
- (*d*) The Far East Rand in Witwatersrand System time:—
  - (i) Conditions of deposition.
  - (ii) Jeppestown Main Bird Unconformity.
  - (iii) Kimberley-Elsburg Series Unconformity.
  - (iv) Igneous Rocks of Pre-Ventersdorp System Age.
  - (v) Summary.
- (*e*) The Far East Rand in Pre-Transvaal Post-Witwatersrand System time:—
  - (i) General discussion.
  - (ii) Thickness of Sediments of Witwatersrand System in relation to extent of Tectonic basining.
  - (iii) Significance of Composition of Igneous Rock associated with the Witwatersrand System.
  - (iv) Significance of various phases of Ventersdorp System.
  - (v) Cause of Basining.
  - (vi) Discussion.

The Old Granite Masses to north-west and south-east of the Far East Rand.

To the north of the Central Rand and north-west of the Far East Rand, roughly between Johannesburg and Pretoria, is a large mass of Old Granite of Pre-Witwatersrand age, and to the south-east of the Far East Rand between Heidelberg and Leslie is another mass of Older Granite. The following are discussions on the possible effects these masses may have had on the structure of the Witwatersrand System in their vicinity.

(*a*) The Johannesburg-Pretoria Old Granite as Portion of Floor of Deposition for Witwatersrand and Ventersdorp Systems.

In the area between Johannesburg and Pretoria is a mass of the Old Granite roughly oval in shape with the long axis trending approximately east and west.

This mass of Old Granite is intrusive into the schists of the Primitive Systems and remnants of these schists, as Willemse states (1933), have an "arcuate arrangement" around the granite.

The southern rim of this Old Granite and remnants of the schists of the Primitive Systems situated around its edges form a portion of the floor on which the sediments of the Witwatersrand System of the Central Rand and north-western portion of the Far East Rand were deposited.

The contact between the Witwatersrand System and the Old Granite and schists as exposed at surface is a fairly regular curve, the curve being concave to the north.

The regularity of the extensive and comparatively thin zone of sediments, seldom more than 600 feet and usually less, forming the base of the Witwatersrand System in this area, namely, the Orange Grove Quartzites, indicates that the floor on which these quartzites were deposited was a very regular plain. In other words the floor of Old Granite and schists on which the Witwatersrand System depositional basin was instituted was in this area a plain.

Thick sedimentary systems are most probably instituted in one of two ways, namely:—

(i) According to A. C. Lawson (1927): "If we start with a high continental margin bordering a shallow sea, the two regions being in isostatic balance, the degradation of land determines a negative load, which will be compensated by the transfer of mass from beneath the sea to beneath the land. The latter will accordingly rise and the sea floor will be depressed. The depression will inaugurate a geosyncline, which thus becomes a trap for sediments and so tends to fill up."

(ii) According to W. H. Bucher depositional basins are instituted in "furrows" and he states: "The origin of 'welts and 'furrows' is independent of the elevation of the earth's crust; it is especially independent of the distinction of the two dominant levels, the sea bottoms and the continental platforms."

Further in defining a geosyncline Bucher considers that: "As a downward fold of the crust, in contrast to folds in sediments, it becomes a synonym of the neutral word "furrow"—"The largest part of the terrigeneous sediments that fill a geosyncline was derived from highlands closely adjoining them, namely, the 'Welt'."

In both the above theories there are highlands, from which the sediments are derived, in close proximity to the depositional basin.

The highlands from which the sediments of the Witwatersrand System were derived were most probably a short distance to the north and north-west of the present outcrop of these beds. The long ridge of the "Older Granites" which trends north-west and south-east, and of which the Johannesburg-Pretoria Old Granite forms the eastern sector, is possibly a portion of this original highland or "welt".

As Lawson has shown (1922), the edge of a depositional basin is a major line of weakness, and Brink (1950) in describing the faulting of the Nuwerus area states:—

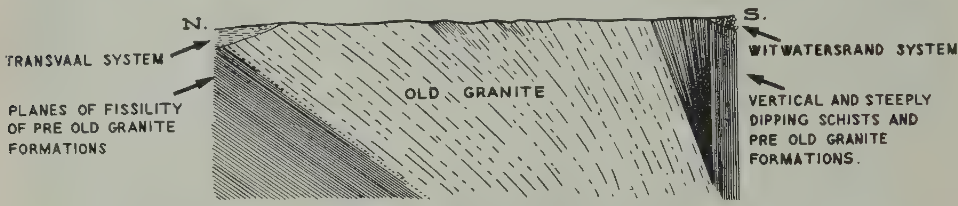
> "While the Karroo geosyncline was sinking in the present Bushmanland, relative uplift occurred along an approximately north-west-south-east axis on its western flank. Between these two areas tensional forces were active which were strongest near the areas of greatest relative uplift or rigidity, causing step-like normal faulting with downthrow towards the geosynclinal side of the present area."

Thus during the deposition of the Witwatersrand System the edge of the depositional basin would become a major line of weakness, and in this region the greatest amount of faulting and movement could be expected.

The mass of the Johannesburg-Pretoria Old Granite as exposed to-day quite probably closely reflects the shape of the original pluton which intruded into the

Primitive Systems. As Willemse (1933) states: "If we assume that the present surface represents an inclined section of the intrusion the arcuate arrangement of the planar structures is accounted for", also "Although absolute conclusive evidence is lacking the form of the granite intrusion appears to be interformational", that is, interformational in Pre-Witwatersrand time.

The periphery of this mass of Old Granite, composed to a large extent of schist of the Primitive Systems, is very likely a zone of weakness (see Fig. 5).



SECTION ACROSS JOHANNESBURG – PRETORIA OLD GRANITE.

(AFTER J. WILLEMSE)

Fig. 5

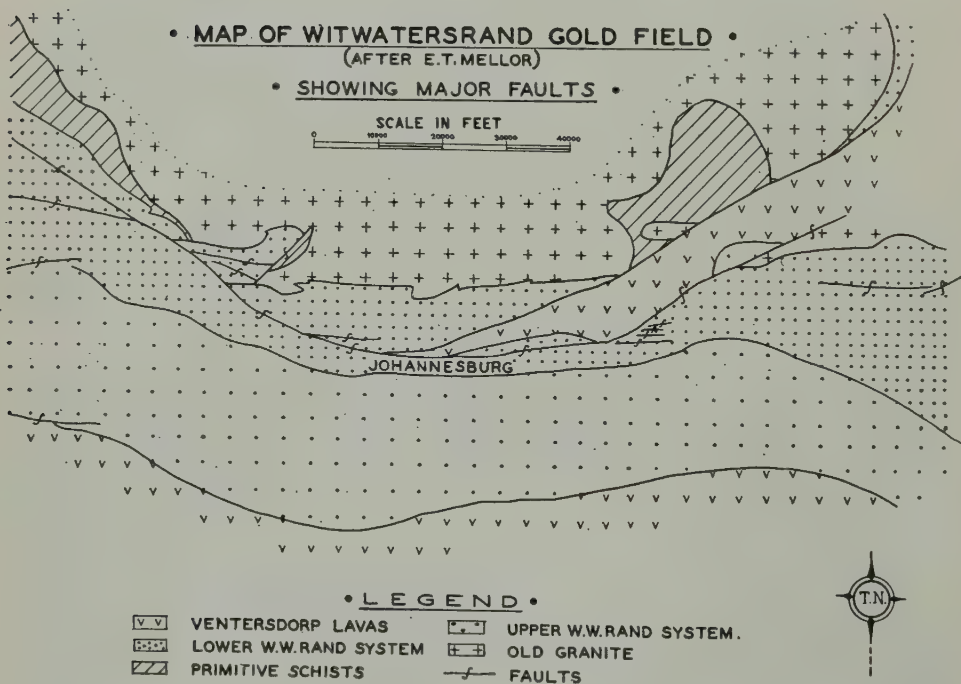


Fig. 6

This sketch clearly indicates the steep "grain" of the schists on the south side of the Old Granite on which the Witwatersrand System rests.

On the Central and north-west of the Far East Rand sediments of the Lower Witwatersrand System, not very competent rocks, have been displaced by large strike faults. The exact dips of these faults are not known except that they are steep and not far off vertical, and the vertical displacements on these faults, although not known accurately, are some thousands of feet (see Fig. 6).

The general trend of these faults is parallel to the strike of the Witwatersrand Sediments, but some, as for instance the one near Rietfontein Consolidated Mines, Limited, curves to the north-east and cuts across the bedding of the sediments and is roughly parallel to the exposed edge of the Old Granite.

In the north-eastern portion of the Central Rand volcanic breccias of the Ventersdorp System are apparently displaced by these large faults, indicating that the faults occurred in post-Ventersdorp time. In other areas the volcanic breccias overlie unconformably the Lower Witwatersrand Sediments and the Old Granite.

On the Central Rand the sediments of the Upper Witwatersrand are remarkably free of large strike faults, although there are a few minor strike faults, with almost vertical dips, near the outcrop of the Main Bird Series along the eastern portion of the Central Rand (Mellor, 1911).

There has been no satisfactory correlation of the sediments of the Rietfontein Consolidated Mines, Limited, although Mellor considered them as the equivalent of the Main Bird Series.

These beds are very similar to those of the known Main Bird Series, although the whole succession of rock types is thinner and there are a number of irregularly bedded conglomerate bands. The high gold content of these conglomerate bands indicates that they are the equivalent of the Main Bird Series.

The sediments of the Rietfontein Consolidated Mines, Limited are probably the equivalent of the Main Bird Series and represent this series deposited near the edge of the depositional basin, which accounts for the fact that these deposits are thinner than those further south and also for the rather irregular bedding.

Summary and Discussion.

The Witwatersrand System was instituted on a plane surface, and the edge of the depositional basin which developed subsequently is a major line of weakness.

The large normal strike faults in the Lower Witwatersrand System along the Central Rand indicate a line of weakness and this line of weakness is probably in close proximity to the edge of the depositional basin.

As the sediments of the Witwatersrand System were probably derived from highlands in close proximity to the depositional basin, and as the sediments of Rietfontein Consolidated Mines Limited probably represent deposits near the edge of the depositional basin, they indicate that the edge of the depositional basin is in close proximity to the present outcrop of the Witwatersrand System on the Central Rand.

The periphery of the mass of the Johannesburg-Pretoria Old Granite, composed to a large extent of schists, is very likely a zone of weakness.

The large normal faults in the Lower Witwatersrand System which strike in a curved line, roughly parallel to the exposed surface of the Old Granite, lie in this peripheral weak zone surrounding the Old Granite.

All the above conclusions indicate that the southern edge of the Johannesburg-Pretoria Old Granite is in close proximity to the edge of the depositional basin, and also account for the two sets of large normal faults in the Lower Witwatersrand System

which in the Central Rand coincide, but which both to the east and west of this area separate, some striking roughly parallel to the strike of the sediments and some parallel to the edge of the Old Granite.

Also as the floor of the Witwatersrand System in this area is composed of the Old Granite and schists of the Primitive System, the Old Granite pluton would be the more stable and probably played a major role in determining the edge of the depositional basin once the basin began to form.

During the tectonic deformation of the region of deposition of the Witwatersrand System into a large basin the Johannesburg-Pretoria Old Granite would again be more stable than the schists.

The relative stability of this mass of Old Granite is thus probably due partly to its proximity to the edge of the depositional basin and partly to its being a pluton, which is relatively more stable than the schists and other masses of Old Granite forming the floor of the Witwatersrand System. It is probable that during the deposition, and during the tectonic basining of the Witwatersrand System, there was a certain amount of sympathetic uplift of the Johannesburg-Pretoria Old Granite so that isostatic equilibrium could be maintained. There is little evidence, however, that this mass of granite was domed up as was the granite of the Parys-Vredefort dome.

(*b*) The Old Granite to the Immediate East and South-east of Nigel.

In the area roughly from five miles east to seven miles south-east of the village of Nigel is a mass of Old Granite, and from the isolated exposures it would appear that the mass may be dome-shaped. The contact between the Old Granite and the overlying Witwatersrand System is exposed at a number of points to the east and south-east of the town of Nigel, but beyond these exposures very little is known of this mass.

These exposures do show, however, that the contact between the two rock groups is a wavy line on surface, the waves being the continuation of those anticlines and synclines found on the Main Reef Horizon (see Fig. 10). The anticlinal fold on the contact of the Old Granite and the Witwatersrand System on the farm Vrisgewaag 337 can be followed across all the sediments of the Lower Witwatersrand into the Upper Witwatersrand. It is the continuation of the anticlinal fold encountered on the Main Reef Horizon on the mining properties of Nigel G.M. Co., Ltd., The Sub Nigel, Limited and Spaarwater G.M. Co., Ltd., and also a continuation of the anticline encountered in the Ventersdorp System on the mining properties of West Vlakfontein G.M. Co., Ltd., Vlakfontein G.M. Co., Ltd., and The Sub Nigel, Limited.

The synclinal fold on the contact of the Old Granite and Witwatersrand System on the farm Uitkyk 97 is the continuation of the syncline on the Main Reef which is present on the mining properties of Nigel G.M. Co., Ltd., Marievale Consolidated Mines, Ltd., and Vogelstruisbult G.M. Areas, Ltd.; and the anticlinal fold on the farm Holgatfontein 127 is a continuation of the Marievale-Vogelstruisbult-Springs anticline on the Main Reef.

This folding in which the folds can be traced through the whole thickness of the sediments of the Witwatersrand System, the lavas of the Ventersdorp System, and to at least the outer edge of the Old Granite, trends from south-south-east to north-north-west and is roughly at right angles to the strike of the contact between the granite and the Witwatersrand System.

The folding could not have been formed by the Old Granite being domed or pushed up, as the axes of the folds all trend from south-south-east to north-north-west

and do not bend round or lie parallel to the edges of the Old Granite as in the Parys-Vredefort area.

In this area neither the Ventersdorp nor the Transvaal System has been found in contact with the Old Granite. On the southern edge of the Old Granite the basic lavas of the Ventersdorp System are separated from it by the large Sugarbush Fault, and the proximity of the outcrop of the Ventersdorp and Old Granite is obviously due to the throw of the Sugarbush Fault, the Old Granite itself being displaced by the fault. The attitudes of the Ventersdorp and Transvaal Systems, although not in contact with the Old Granite, show no sign of being domed by it. Near the contact of the Old Granite and the Witwatersrand System are a few minor strike faults with slight duplication of some of the sediments of the Lower Witwatersrand System (Rogers, 1922). This duplication, due to overcreep or reverse faulting, suggests relative downward movement of the Old Granite with respect to the Witwatersrand System; also the rock formations have been folded, and many, possibly all of the faults found, including those duplicating the sediments, are related to the folding, which has therefore not been caused by upward movement of the Old Granite.

From the lack of any structures associated with doming, it would appear that the Old Granite to the east of Nigel has not been domed but rather has been a relatively stable mass during the basining of the Witwatersrand System. The curved line of the exposures of the Old Granite at its contact with the Witwatersrand System is related to the folding and not to any doming of the Old Granite.

(*c*) The Pre-Witwatersrand System Floor.

The Witwatersrand System was deposited on a floor composed of various masses of the Old Granite and in isolated instances on remnants of the schists of the Primitive System. The cleavage of these schists of the Primitive System in the Heidelberg area strikes west-north-west and dips at very high angles (Rogers), and in this area the Old Granite is gneissic in many places and the trend of this gneissose structure also is west-north-west (Rogers).

Willemse (1933) in his examination of the mass of Old Granite which lies between Johannesburg and Pretoria found numerous shear zones but no regional "grain" in the granite, and the crush zones and schistosity encountered he considers are related to movement after and possibly during the deposition of the Witwatersrand.

The "grain" of the pre-Witwatersrand floor in the vicinity of the Central and Far East Rand thus does not help in determining whether the "grain" affected the orientation of the depositional basin.

In the Lowveld of the Transvaal Province, especially around Barberton and the Murchison Range near Leydsdorp, are considerable masses of schists of the Primitive Systems, and the regional grain of these schists is from east-north-east to west-south-west, and in places from north-east to south-west.

In this connection the distribution of known dyke swarms in the north-eastern Transvaal is also of considerable interest.

To the north of Leydsdorp there are approximately 600 dolerite dykes which trend north-east and south-west and only a few follow other directions.

To the north of Piet Retief are about 100 dykes which trend north-west and south-east and 20 dykes which trend north-east and south-west.

From the distribution of the dolerite dyke swarms in the North-eastern Transvaal a north-east to south-west direction is inferred which suggests an inherent direction of weakness in the Basement Complex.

It is tentatively suggested that the "grain" of this Basement Complex in the Lowveld of the North-eastern Transvaal as indicated by the dyke swarms in this area affected the orientation of the mobile belt in which the Witwatersrand deposition basin became instituted.

(*d*) The Far East Rand in Witwatersrand System Time.

(i) *Conditions of Deposition.*

The sediments of the Witwatersrand System are remarkably conformable for a series of sediments of this thickness (on the Central Rand the System is approximately 25,000 feet thick). It is possible to trace individual comparatively thin horizons over the whole known area of the Witwatersrand System and this fact has facilitated correlation of the various series in widely separated areas. As Nel states: "Characteristic of the Witwatersrand System is the presence in it of some remarkably persistent rock units that retain their distinctive character over great distances, and as several of these well-known key horizons or "markers" of the Rand are also present in the Klerksdorp-Ventersdorp area, the correlation and classification of the strata exposed was considerably facilitated".

Rogers found that "The major sub-divisions of the Witwatersrand System established by Dr. Mellor on the Rand were found to be clearly recognisable in Heidelberg and so were many of the minor sub-divisions." Nel found that the same conditions applied to the area between Potchefstroom and Klerksdorp.

There is, however, considerable thinning of the sediments in certain directions, namely, to the east and south-east (Mellor, 1915 and Jones, 1936), but individual horizons retain their characteristics and the following extract from Rogers (1922) gives the extent of this thinning:—

> "The foregoing account of the Witwatersrand beds in Heidelberg shows that the thickness of the beds is much less than it is on the Rand while the thickness of the System in Heidelberg is 57 per cent of what it is on the Rand. The Upper Witwatersrand beds in Heidelberg are only about 52 per cent of their thickness on the Rand, and the Jeppestown, Government Reef Series and Hospital Hill Series in Heidelberg are respectively 52 per cent, 62 per cent and 66 per cent of their thickness on the Rand."

The presence of thin, extensive and lithologically persistent horizons at repeated intervals in the stratigraphic column of both the Lower and Upper Witwatersrand Systems, indicates that the sediments were deposited on almost horizontal surfaces, and that the depth of water under which these sediments were deposited was sufficiently shallow for wave action to spread them over large areas.

Sharpe (1949) in a recent paper, to explain the various phases of the Witwatersrand System, considers that these phases or cycles of sedimentation are due to uplift and subsidence. The sediments were certainly deposited under water of varying depth, but whether, apart from minor unconformities, these cycles were caused by upward movement is doubtful. As Haughton in discussing Sharpe's paper states: "the area of deposition must in general have been subsiding, and the rate of subsidence must have been equal to, greater than, or less than the rate at which material was being deposited", and further that the variability shown by the beds can be explained as "the result of ryhthmic rates of subsidence in the floor of the sinking basin and would not require periodic movement of uplift and subsidence."

The regularity of the system as a whole indicates that subsidence and deposition were closely related, and it is possible that the slow sinking of the crust was due to the

load of sediments with adjustment of the crust so that isostatic equilibrium could be maintained.

Whether the slow sinking of the crust to allow for the deposition of, in places up to 25,000 feet of sediments, was due to isostatic compensation, or due to what Bucher calls up and down movements in the earth's crust in certain mobile belts, the fact remains that the floor of the Witwatersrand System was depressed up to depths of 25,000 feet in relation to the area to the north and north-west during the time in which the Witwatersrand System was being deposited.

The regularity of the stratigraphic column at the outcrop indicates that only those sediments that were deposited on almost horizontal surfaces remain, and that those sediments near the edge of the original depositional basin have been eroded away. The attitudes of the sediments on the Central, East and Far East Rand are not the attitudes of the sediments as deposited, and are due to subsequent movements. See Fig. 7.

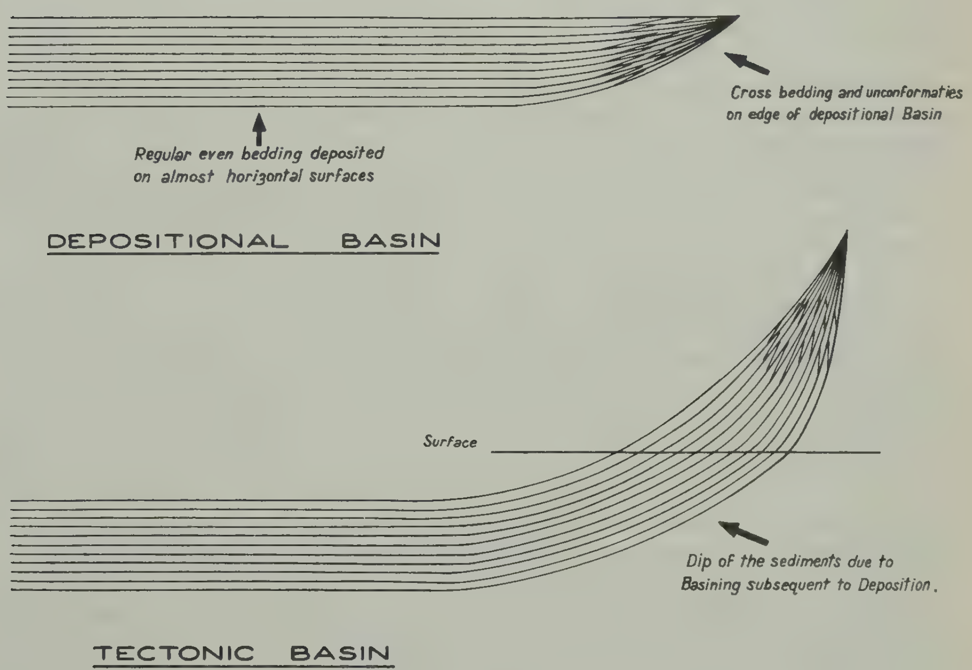


Fig. 7

Although on the whole the sediments and series of the Witwatersrand System are conformable there are some horizons which display a certain degree of unconformity. In the Klerksdorp area Nel (1939) has shown that there is a considerable unconformity in the Upper Witwatersrand System, also on the Central, East and Far East Rand sectors of the Witwatersrand there are unconformities, and the following are descriptions of two such unconformities in the Witwatersrand System in the latter areas.

(ii) *Jeppestown-Main Bird Series Unconformity.*

On the Central, West and Far West Rand sectors of the Witwatersrand there are some 200 feet of quartzites and grits between the Jeppestown Shales and the Main Reef. These quartzites are not present on the Far East Rand, except possibly in some erosion channels.

On the East Rand Proprietary Mines, Ltd., which is situated to the immediate west of the Far East Rand, this horizon of quartzites gradually thins out till on the eastern boundary of the mine the "Main Reef" rests directly on the Jeppestown Shales. This gradual thinning towards the east is due to non-deposition of the quartzites and not to erosion. (Carleton Jones, 1936.)

This disconformity was probably caused by uneven subsidence of the depositional basin in which the sediments of the Witwatersrand System were being deposited. According to recent definitions this type of unconformity is called a static disconformity (Grabau), or disconformity due to non-deposition (Pettijohn).

In addition to this disconformity, as stated previously, there was uplift folding and erosion in certain areas on the Far East Rand prior to the deposition of the Main Reef and Main Bird Series.

This folding and erosion in the pre-Main Reef surface was on a relatively small scale, the amplitude of the folds seldom exceeding 200 feet, and the erosion channels seldom being more than 200 feet deep. Not sufficient work has been done in the foot-wall of the Main Reef to determine the depth or trend of this local folding, and also this folding has been found only in certain areas, and over the greater portion of the Far East Rand is not present; but in those areas where it is present the Main Reef is definitely unconformable to the underlying sediments.

It is significant that this puckering and erosion occurs in that portion of the Witwatersrand where the System as a whole is considerably thinner than normally, but there is insufficient evidence to demonstrate the cause of this minor unconformity, and to decide whether it was due to tectonic action or to the possible uneven subsidence in the depositional basin in which the sediments of the Witwatersrand System were being deposited, accompanied by slumping and folding of the upper portion of the Jeppestown Series.

(iii) *Kimberley Elsburg Series Unconformity.*

The May Reef of the Kimberley-Elsburg Series rests unconformably on the underlying sediments in portions of the Far East Rand, and probably in other areas as well.

De Jager (1949) in a discussion on Sharpe's paper gives details of the May Reef and its relation to the footwall and hanging wall sediments. In addition he has kindly consented to my using some results of his as yet unpublished report. He has established the following facts. After the laying down of the Kimberley Shales and also a few hundred feet of quartzites and conglomerates above these shales, there was a pause in the deposition of the sediments of the Upper Witwatersrand System. During this pause there was slight puckering and folding of the then surface of the Witwatersrand System, and these puckers and folds have crests trending from north-west to south-east. These folds are very gentle and the amplitude seldom more than 200 feet, i.e., the vertical height from crest to trough. After, and possibly during this folding, there was a period of erosion and the surface was again reduced to a plane. In isolated places the erosion was comparatively deep and all sediments above and including the Kimberley Shales were eroded away.

The May Reef is a quartzitic conglomerate resting on this planed surface, and its high gold content in certain places is related to its footwall and is very likely due to a reconcentration of the gold which was present in small quantities in the Kimberley Reefs prior to these being folded and eroded.

This intraformational deformation is very local, the sediments a few hundred feet below this line of unconformity being parallel to those above it, and although very likely present in other parts of the Witwatersrand, in most places there is no noticeable break in the deposition of the Kimberley Elsburg Series. (See Plate III, Section V.)

These two local unconformities of low degree, namely, the Main Reef and May Reef unconformities, are of great economic importance to individual mines and warrant a great deal of detailed examination, but become less significant when considering the deposition of the Witwatersrand System as a whole.

(iv) *Igneous Rocks of Pre-Ventersdorp System Age.*

There are two known relatively thin basic lava zones in the Witwatersrand System, namely, the Jeppestown Amygdaloidal Lavas and the Bird Amygdaloidal Lava.

The Jeppestown Amygdaloidal Lavas are seldom more than 400 feet thick and usualy very much less, and form a remarkably persistent horizon situated roughly in the centre of the Jeppestown Series. They have been found wherever this series has been probed, and serve as a very useful marker, having been found on the Far East Rand, the Far West Rand and as far south as the Odendaalsrus Gold Fields.

The Bird Amygdaloidal Lavas are not as persistent as those of the Jeppestown, but are present over the whole of the Far East Rand, and at least a tuffaceous phase of this extrusive is present over the whole length of the Central Rand and possibly portions of the Far West Rand. The Bird Amygdaloidal Lavas are seldom more than 300 feet thick and often the lavas of individual flows are separated by thin sedimentary bands which, however, are not very persistent. These lavas are situated in the upper half of the Main Bird Series.

These two extrusive igneous phases in the Witwatersrand System are completely conformable to the sediments above and below them. Except for some acid lavas in the Klerksdorp sector, which in that area form the base of the System, no acid lavas are found in the Witwatersrand System.

In all underground work on the Witwatersrand numerous intrusions, both dykes and sills, are encountered, and almost invariably these are basic in composition. In some cases it is possible to ascertain the relative ages of these intrusions such as those of Karroo age and Transvaal age, but there has been no really satisfactory study of the intrusions of pre-Transvaal age. The intrusives that accompanied the Jeppestown extrusive naturally would not be encountered in the younger Main Bird Series, and thus would not be found in mine workings, but intrusives associated with the Bird Amygdaloidal Lavas must be present. Unfortunately no distinction between these and later basic intrusives has been made.

These two basic extrusives indicate that during the deposition of the Witwatersrand System there must have been a little cracking of the floor of the depositional basin to relieve tension, and that through these fissures a certain amount of igneous material welled up from below, portions of which reached the top of the sediments and spread out over them as lavas.

This foundering and cracking of the floor must have been on a relatively small scale and was insufficient to have any effect on the attitude of the sediments at the top

of the succession, as the sediments above the lavas are conformable to those below them.

(v) *Summary.*

Apart from the gradual thinning of the sediments of the Witwatersrand System to the east and south-east, the above-mentioned unconformities of low degree and the two minor igneous extrusives, and possibly other deformations and interruptions in the deposition of the sediments of a similar nature, the sediments of the Witwatersrand System on the Central, East and Far East Rand sectors were deposited on plane surfaces, and there are no signs of sporadic or regional synclinal subsidence during their deposition, nor is there any sign that the land mass to the north or north-east was domed up.

The series of the Witwatersrand System on the Far East Rand were deposited on plains which are conformable to each other, and there is no sign of regional tectonic basining during their deposition.

The deformation of the Witwatersrand System on the Far East Rand did not commence during the deposition of these beds.

(*e*) THE FAR EAST RAND IN PRE-TRANSVAAL POST-WITWATERSRAND SYSTEM TIME.

(i) *General Discussion.*

On the Central, East and Far East Rand sectors of the Witwatersrand, the deposition of the sediments of the Witwatersrand System was apparently stopped by the pouring out of the lavas of the Ventersdorp System.

In no portion of the Far East, east or south of the Central Rand do the lavas of the Ventersdorp System rest unconformably on the underlying sediments of the Witwatersrand System.

As shown in sub-division (*d*) of this chapter there is no evidence of regional tectonic deformation during the time of deposition of the Witwatersrand sediments on the Far East Rand.

The lavas of the Ventersdorp System are folded and dip in exactly the same manner as the underlying sediments of the Witwatersrand System in the Central, East and Far East Rand sectors of the Witwatersrand.

After the pouring out of the lavas there was a long pause before any further sediments were deposited, and it was during this pause that the major deformation of the Witwatersrand System took place, and the sediments of the Witwatersrand System and lavas of the Ventersdorp System assumed their present basin-like shape. The surface was then reduced to approximately a plain and the Transvaal System deposited unconformably on this surface composed of either the Old Granite, Witwatersrand or Ventersdorp Systems.

In the Central, East and Far East Rand sectors of the Witwatersrand the tectonic deformation of the Witwatersrand System took place in post-Ventersdorp time, or rather post- that phase of the Ventersdorp System present in these sectors.

(ii) *Thickness of Sediments of Witwatersrand System in relation to extent of Tectonic Basining.*

As described previously the sediments of the Witwatersrand System on the Far East Rand generally dip to the south-south-east from the northern edge of the basin, west-south-west from the eastern edge and north-north-west from the southern edge. Although there is no western edge of the basin, the western rim is roughly on

a line from the western portion of West Vlakfontein G.M. Co., Ltd. to the western portion of Van Dyk Consolidated Mines, Limited, and west of this line the beds dip at relatively steep angles into the Central Rand Basin.

From section lines drawn across the Far East Rand from north-north-west to south-south-east it is noticed that the deepest portion of the Main Reef is roughly midway along these section lines, that the curves formed by the Main Reef are very regular, and that there is little or no sign of folding parallel to the north-north-western or south-south-eastern rim of the basin.

The sections show that the Main Reef has the greatest amount of curve in the western portion of the Far East Rand, the difference in elevation of the Main Reef between outcrop or sub-outcrop and the centre of the basin being at least 8,000 feet. Sections parallel and to the east show a gradual decrease in the curve until, near the Main Reef sub-outcrop along the eastern fringe of the basin, the curve almost disappears and is nearly a straight line.

Section lines drawn across the Far East Rand in directions other than from north-north-west to south-south-east, do not show anything like the same regularity, and this is due to a number of north-north-west to south-south-east trending folds, but these sections do show that the basining of sediments decreases towards the eastern fringe of the Far East Rand Basin.

On the Far East Rand the basining of the sediments of the Witwatersrand System is greatest in the western portion where the sediments are thickest, and gradually decreases in an easterly direction.

To the west of the Far East Rand, that is over the East and Central Rand, the basining is considerably greater and the difference in elevation on the Main Reef horizon from outcrop to the bottom of the basin is probably of the order of 20,000 feet. On the Central Rand the sediments are almost twice as thick as those on the Far East Rand.

The tectonic basining of the sediments of the Witwatersrand is clearly related to the total thickness of the sediments.

(iii) *Significance of Composition of Igneous Rocks associated with the Sediments of the Witwatersrand System.*

There are three igneous extrusive periods closely associated with the deposition of the sediments of the Witwatersrand System, namely the Jeppestown and Bird Amygdaloidal Lavas and the Ventersdorp Lavas. All these extrusive rocks are basic in composition.

In underground work on the Witwatersrand mines numerous intrusives, both dykes and sills, are encountered and almost invariably these are basic in composition.

Although there is no satisfactory correlation of the igneous intrusions of pre-Transvaal Age many of these basic dykes and sills are most probably related to the three extrusive phases mentioned above.

In the mine workings large sills are encountered which often transgress from the footwall to the hanging-wall and back again into the footwall of the Main Reef, and where they transgress they displace the Main Reef. These displacements are usually of the order of 200 feet, but are often as great as 400 feet, depending on the thickness of the sill. Large sills also are encountered near the horizon of the Kimberley Shales, and these transgress from below, to into, to above the shales. All these sills simply part the sediments and there is no apparent absorption of the sedimentary rocks. The sills are basic in composition and are most probably intrusive phases of the Ventersdorp extrusives (Ellis, 1946).

The lines where these sills cut across the Main Reef and turn into dykes are on the whole very irregular.

In addition to these intrusives in the Upper Witwatersrand, numerous igneous intrusives occur in the Lower Witwatersrand and these are too almost invariably basic in composition.

One aspect of these igneous rocks that helps to solve the problem of the deformation of the Witwatersrand System is the fact that they are basic, and as basic and not acid rocks are normally associated with "furrows" (Bucher) or areas that are subsiding, they indicate that subsidence continued to at least post-Ventersdorp time.

Although a few acid dykes are found on the Far East Rand these, as Rogers has shown, are of post-Ventersdorp pre-Karroo System age. Also the granophyres that occur on Marievale Consolidated Mines, Ltd., Vogelstruisbult G.M. Areas, Ltd., The Sub Nigel, Limited and Vlakfontein G.M. Co., Ltd. occupy fissures that are related to the folds of the Far East Rand.

These acid dykes and intrusives indicate uplift, and as has been shown this uplift and folding occurred in post-Ventersdorp time and as will be shown later in the discussion the folding and associated faults are of pre-Transvaal age.

Significance of various Phases of the Ventersdorp System.

The Ventersdorp lavas that overlie the Upper Witwatersrand System are basic in composition.

One rather significant fact is that there are acid phases of the Ventersdorp System in areas outside the synclinal basin of the Witwatersrand System. In the vicinity of Klerksdorp there are some thin flows of acid lava overlying the Witwatersrand System, but in this there was uplift prior to the laying down of the lavas. Nel (1939) states: "In the Klerksdorp region the Ventersdorp formation indicates an uplift and abrasion of the underlying formation before its first members were laid down." Apart from this area the author knows of no instances where large flows of acid lava overlie the sediments of the Witwatersrand System.

This is very clearly shown in investigations to the west and north-west of Odendaalsrus, and also to the west of Klerksdorp, where the Ventersdorp System is composed of alternating layers of acid and basic lavas with some sedimentary zones, but in these parts the acid lavas predominate and are very widespread, covering an area of well over 2,000 square miles. No sediments of the Witwatersrand System have been discovered beneath these acid lavas.

Acid magmas have a lower melting point than basic magmas, and acid magmas flow at considerably lower temperatures than basic magmas (Greig a.o., 1929); also acid rocks have a considerably lower specific gravity than basic rocks.

These facts indicate that during the deformation of the earth's crust the acid magmas would collect under the rising portion and basic magmas under the sinking portion of the deformed crust, and that if conditions were suitable these magmas would break through and flow.

It is very probable that the acid lavas of the Ventersdorp System related to those portions of the Orange Free State and Western Transvaal that were rising, while the basic lavas of the Ventersdorp System are related to those regions where the Witwatersrand was being deposited, namely, in a subsiding area.

In the author's opinion no sediments of the Upper Witwatersrand System will be found below the acid lavas to the west of Odendaalsrus and Bothaville.

From the numerous boreholes drilled in the Orange Free State it has been established that there are at least two phases of the Ventersdorp System in which basic

lavas were poured out. The first basic lava phase overlies the Witwatersrand Sediments with apparent conformity; as Borchers (1951) states, "The Ventersdorp Lower Lava is seen to rest with regional conformity on the Upper Witwatersrand". After this extrusive phase the sediments and lavas were deformed and eroded, and in most places sediments were deposited unconformably on this eroded surface. These sediments are composed largely of rounded and angular pebbles of volcanic material but also of quartzites and shales. After this sedimentary phase further basic lavas were poured out and overlie unconformably either the sediments or earlier lavas or both. As Borchers explains: "There is no known evidence of faulting action during the period of effusion of the Lower Ventersdorp Lavas, but immediately thereafter, and before the commencement of the Upper Ventersdorp phase, a considerable period of colossal faulting and peneplanation must have taken place."

In the area a few miles to the south and south-east of Odendaalsrus the Witwatersrand System and the first or lower extrusive phase of the Ventersdorp System (conformable to the Witwatersrand System and comprising about 2,000 feet of basic lavas) foundered and was deformed into a synclinal basin. The Witwatersrand System and these overlying lavas dip at approximately 30° to the east. After this foundering there was considerable erosion of the edge of the basin and then further sediments and the Upper Lavas of the Ventersdorp System were deposited on this eroded surface. Subsequent to the latter extrusives there was further subsidence, but the base of the Upper Lavas seldom dips at more than 5° so that this later subsidence was on a very much smaller scale. The Karroo System overlies the Witwatersrand and Ventersdorp Systems unconformably in the Odendaalsrus area and these beds are almost horizontal.

In the area to the immediate south and south-east of Odendaalsrus the foundering and major basining of the Witwatersrand System took place during early Ventersdorp System time and continued to a slight degree in post-Ventersdorp time.

On the Far West Rand the lavas overlie the Witwatersrand Sediments unconformably and although the angle of unconformity is small, seldom more than 4°, the unconformity is quite definite.

On Venterspost G.M. Co., Ltd. the Ventersdorp Contact Reef dips at about 35° to the east and the Main Reef at about 39° to the east.

The fact that the Ventersdorp Contact Reef is the base of the Ventersdorp Lavas indicates that the major basining of the Witwatersrand and Ventersdorp Systems took place after the pouring out of the lavas of the Ventersdorp System.

The igneous rocks both intrusive and extrusive associated with the Witwatersrand System, indicate that subsidence of the Witwatersrand depositional basin continued during Ventersdorp time. The lavas of the Ventersdorp System in the Odendaalsrus sector indicate that the major basining of the Witwatersrand System took place during Ventersdorp time, but in all other areas except around Klerksdorp and north of the Central Rand major basining appears to have occurred in post-Ventersdorp time.

The following is, however, another aspect which should be considered in fixing the time of the major basining of the Witwatersrand System:

It is generally accepted that there was a considerable pause between the deposition of the sediments of the Witwatersrand System and the pouring out of the lavas of the Ventersdorp System, and this has been based on the fact that in places the lavas rest unconformably on the Witwatersrand System. The conformity of these two systems on the Central, East and Far East Rand, in the Vredefort area and in the Odendaalsrus Gold Fields, and the unconformable relation between the later basic lavas and sediments of the Ventersdorp System and the earlier lavas of the Venters-

dorp and Witwatersrand Systems in the Odendaalsrus area, indicate that the long pause in time may have occurred during what is at present called Ventersdorp time.

A simple explanation is that those lavas that are conformable to the underlying Witwatersrand System should be classed as an extrusive igneous phase of the Upper Witwatersrand System, and that those lavas that overlie the lavas and sediments unconformably are of Ventersdorp System age.

In other words the Ventersdorp System as understood to-day should be subdivided and some of the basic lavas classed as belonging to the Witwatersrand System and others as belonging to the Ventersdorp System.

This would mean that the deposition of the sediments of the Witwatersrand System ceased on the outpouring of basic lavas, and that after this extrusive igneous phase of the Witwatersrand System, which could be called the Elsburg Lavas, both the sediments and the lavas were deformed into a synclinal basin by the more complete or more intense foundering of the floor of the depositional basin. After this basin foundering there was the long pause during which there was considerable peripheral erosion and then only were the lavas of the "Ventersdorp" System poured out. After this latter extrusive phase further foundering occurred but on a smaller scale.

There was slight peripheral deformation in the Far West Rand prior to the pouring out of the "Elsburg" lavas, the angle of unconformity between the lavas and the sediments seldom being more than 4°. The major deformation in this area, however, took place after the pouring out of the lavas. In this area the Elsburg Lavas are unconformable to the underlying sediments, but it is tentatively suggested that this is a local unconformity similar to other unconformities present in the Witwatersrand System.

In the Klerksdorp area there was major deformation prior to the pouring out of the lavas and here it would appear that the major pause in time occurred in pre-Ventersdorp time.

Although there is no evidence of an earlier phase of extrusives in the Klerksdorp area overlying conformably the sediments of the Witwatersrand System, it is possible that such a phase of extrusives exists as shown in Fig. 8.

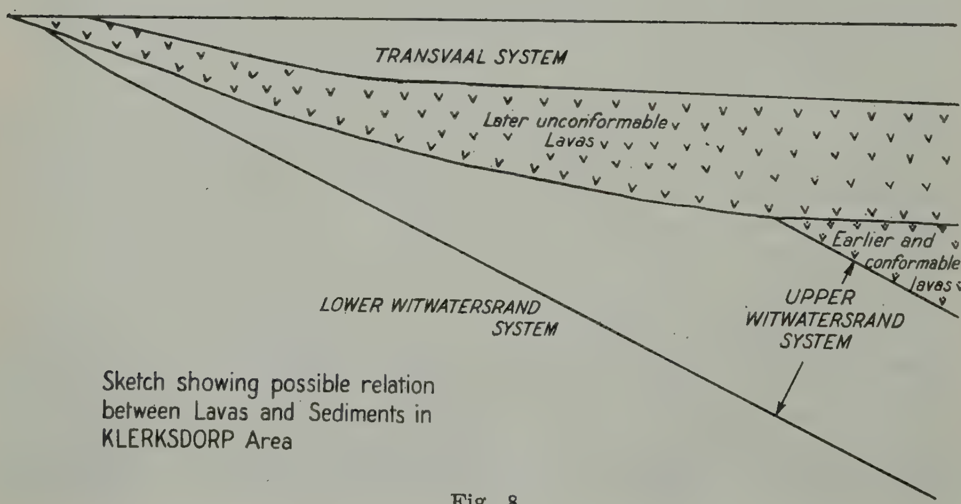


Fig. 8

A relatively thin zone of porphyritic lava is present a few hundred feet above the base of the lavas on the Central and Far East Rand, also in Vredefort, the Far West Rand and Klerksdorp, and as de Kock (1940) states "must be regarded as having been extruded contemporaneously over this entire area".

If this porphyritic phase of a lava flow is in fact a persistent horizon over this large area it is definite evidence of regional unconformity between the Witwatersrand and Ventersdorp Systems, but porphyritic phases of the lavas are not present in the Odendaalsrus area (Borchers, 1951), and although porphyritic phases of lavas are present in the Potchefstroom-Klerksdorp area (Nel, 1939) these occur at various horizons but usually in the lower portions of the Ventersdorp System. On the Far East Rand in No. 3 Shaft at Vogelstruisbult G.M. Area, Ltd. porphyritic phases of the lavas occur at the following depths below the collar of the shaft: between 540 and 660 feet and between 815 and 1,070 feet. The base of the lavas is at 1,287 feet. It is tentatively suggested that such porphyritic horizons may be found to occur at different horizons in other parts of the Witwatersrand basin and that the horizon is not persistent over the whole area.

If this porphyritic phase is not a persistent horizon the condition shown in Fig. 8 is possible.

The above is completely theoretical, but as two ages of extrusives explain so many of the structures, as for instance the unconformity between the lava breccias and sediments in the Bezuidenhout Valley, the conformable relation between the sediments and lavas on the Central Rand, the unconformable relationship between the Witwatersrand and Ventersdorp Systems in the Klerksdorp area, and the apparent conformable relationship in the Vredefort and Odendaalsrus areas, this theory is suggested which is based mainly on the occurrence of two ages of basic extrusives in the relatively small sector of the Witwatersrand, namely, the Odendaalsrus area (Plate I).

If there are in fact two ages of extrusives, the one of Witwatersrand System age and the other of Ventersdorp System age, then the major foundering of the Witwatersrand System into the large synclinal basin occurred in post-Witwatersrand pre-Ventersdorp time, and not during Ventersdorp time. The deformation continued to a lesser degree, however, in Ventersdorp time and also in post-Transvaal System time.

This relegation of the basic lavas to two different systems does not affect the reasoning that the synclinal basining of the sediments was due to subsidence, but is of great importance in determining the time of the deformation.

To avoid confusion in this paper, however, the lavas that overlie the sediments of the Witwatersrand System on the Central and Far East Rand have been, and will be, called by their generally accepted name, lavas of the Ventersdorp System.

Cause of Basining of Witwatersrand System.

The two intraformational igneous extrusive phases, namely the Jeppestown and Bird Amygdaloidal Lavas, indicate that there was a certain amount of release of tension in the floor of the Witwatersrand System during the deposition of the sediments.

By the time that the full succession of sediments, which on the Central Rand amount to nearly 25,000 feet, were deposited it is probable that the elastic limit of the floor of the depositional basin was transgressed causing fractures. With this fracturing of the floor basic magmas again welled up and the lavas of the Ventersdorp System

poured out over the as yet undisturbed uppermost sediments of the Witwatersrand System, undisturbed at least in the Central and Far East Rand. The additional weight of these lavas probably enhanced the shape and extent of the basined Witwatersrand deposits, and the whole sedimentary formation and the overlying lavas slumped. With the slumping are associated the large strike faults in the Lower Witwatersrand System on the Central Rand. Subsequent to the slumping there was probably a further welling up of igneous matter, portions of which may have come up in the opening formed by the large tensional strike faults along the Central Rand. It is possible that this second extrusive phase poured out over the earlier lavas, but there is no evidence of this on the Central or Far East Rand.

DISCUSSION.

The regional shape of the Upper Witwatersrand System is a large roughly oval-shaped synclinal basin, with the long axis a little over 200 miles long and the short axis approximately 100 miles (see Plate I). The regional shape of the Lower Witwatersrand is approximately the same as that of the Upper Witwatersrand and naturally larger. Although too small to be classed as a geosyncline, the Witwatersrand synclinal basin is in many respects similar to one.

A geosyncline is usually defined as a long and relatively narrow, slowly progressing, downward fold in the earth's crust in which thick sedimentary formations were deposited, the downward movement ending in a period of deformation of these sediments by compression causing uplift, thrust faulting and folding (Bucher, 1933, Grabau, 1932, Pettijohn, 1948).

The thickness of the sediments of the Witwatersrand System, up to 25,000 feet, is sufficient for it to be classed as a geosynclinal deposit, but the length and breadth are considerably smaller than required, and although of Pre-Cambrian age, there was not the usual orogenic epoch with folding and uplift, which is invariably associated with geosynclines in other parts of the world.

The depositional basin in which the Witwatersrand System was instituted was possibly situated along a zone of weakness in the Older Granite and Primitive System, and the basin was either what Bucher calls a "furrow" which formed in one of the mobile belts of the earth during crustal tension, and/or was due to the slow sinking of the earth's crust under a load of sediments.

After the deposition of the sediments of the Witwatersrand System, in certain sectors there was foundering in the floor of the depositional basin, namely in the vicinity of Klerksdorp and the Far West Rand. After this foundering there was peripheral erosion of the Witwatersrand sediments, prior to the pouring out of the Ventersdorp Lavas in these sectors. These conditions do not apply over the major portion of the Witwatersrand System, and in the Central, East, Far East Rand, Vredefort and Odendaalsrus sectors of the Witwatersrand there is no sign of foundering and erosion prior to the pouring out of the lavas. In these sectors the lavas of the Ventersdorp System (Elsburg lavas) were poured out on an undisturbed surface of the Elsburg Series of the Witwatersrand System. After, or possibly during, this igneous extrusive phase the Witwatersrand System was deformed into a large synclinal basin. This basining was most probably caused by the elastic limit of the floor of the depositional basin being transgressed, and with subsequent fracturing, the lavas welled up and poured out over locally undisturbed surfaces of the Witwatersrand sediments. After the outpouring of the lavas the floor foundered, and possibly owing to isostatic compensation, both the sediments of the Witwatersrand and the lavas of the Ventersdorp System subsided and formed the synclinal tectonic basin

of very nearly the same shape as at the present time. The shallow basins of the Transvaal System indicate that subsidence continued in post-Transvaal time and this subsidence must have accentuated the basin formed by the foundering of the floor of the Witwatersrand depositional basin. This basining of the Transvaal System is, however, of relatively small degree. See Plate III, Sections III and IV.

VI THE FOLDING OF THE WITWATERSRAND SEDIMENTS ON THE FAR EAST RAND.

The sediments of the Witwatersrand System on the Far East Rand have been folded, and comparable folding is not known in other parts of the synclinal basin of the Witwatersrand System. There are signs, however, of very gentle large-scale warping of the Witwatersrand System, for example the sediments of the West Rand could be considered as being curved to form a very large syncline and the gentle curve of the strike of the sedimentary horizons on the Far West Rand, with the apex near Bank Station, could be considered as an anticlinal bend.

All these gentle curves are on a very large scale, being of the order of 50 miles from crest to crest, and certainly not comparable with the size of the folds of the Far East Rand which measure only a few miles from crest to crest. There is a possibility, however, that these large-scale warps are related to the folds of the Far East Rand.

To arrive at a possible cause of the folding of the Witwatersrand System on the Far East Rand it is necessary to determine the relationship between the basining and folding, also the various types of movement that could produce folding of the kind present on the Far East Rand.

Discussions of these various aspects will be given under the following headings.

(*a*) Relationship between Old Granite and Folds.

(*b*) Other possible causes of Folds:

  (i) Gravity Folds;

  (ii) Convection current Folds.

(*c*) General description of Folds and probable causes.

(*d*) General description of some major Faults and probable causes.

(*e*) The Sugarbush Fault.

(*f*) Summary and Discussion.

(*a*) Relationship between Old Granite and Folds.

The axes of the folds on the Far East Rand strike roughly north-north-west and south-south-east and this is the direction in which the Pretoria-Johannesburg and Nigel Old Granite masses lie relative to each other, the Far East Rand being situated roughly midway between these two masses.

It has been shown that although the major deformation of the Witwatersrand System into a synclinal basin was probably caused by foundering of the floor and subsequent subsidence in post-Ventersdorp pre-Transvaal time, there is evidence of some movement of the Johannesburg-Pretoria Granite mass during and after Ventersdorp time. If this granite mass took part in a horizontal movement towards the Nigel Old Granite mass, either one or both masses moving, such movement would cause folding at right-angles to the direction of movement. The folds of the Far East Rand are parallel to this direction and could not have been caused by the granite masses moving towards each other, or the one mass towards the other.

Again upward movement of these granite masses should cause lateral pressure, but as before no folds on the Far East Rand can be ascribed to pressure from these directions.

Thus although there may have been upward movement of one or both of these granite masses, it is most unlikely that such movement caused the folds in the Witwatersrand System on the Far East Rand.

(*b*) Other Possible Causes of Folds.

(i) "*Gravity Folds*".

Folding is sometimes due to the sediments puckering around the periphery of a depressed or subsiding mass, and is usually referred to as "gravity folding".

In the base of the Witwatersrand System on the Far East Rand the folds were formed after the pouring out of the Ventersdorp Lavas. The sediments of the Witwatersrand were deposited in a depositional basin and both the sediments and the overlying lavas were depressed or subsided into a tectonic synclinal basin. The folds of the Far East Rand are not however peripheral, there being no folds parallel to either the north-north-eastern or south-south-western flanks of the basin, and although in places the folds are parallel to the eastern rim, this is so only over a short distance; the rim of the basin as a whole being curved and the axes of the folds being straight lines.

It is most unlikely that the folds of the Far East Rand were due to "gravity" or rather puckering round the edge of the subsiding mass of sediments of the Witwatersrand System.

The folds of the Witwatersrand System on the Far East Rand could have been caused by the subsiding of a sedimentary system to the east or north-east of the Far East Rand, in other words these folds could be gravity folds near the edge of a sedimentary formation other than the Witwatersrand System.

Boreholes drilled to the east of the Far East Rand have penetrated the Transvaal System and encountered the Old Granite and further east in the vicinity of Ermelo the Karroo System rests on the Old Granite. There is no evidence yet that any system of post-Ventersdorp pre-Transvaal age is present to the east or north-east of the Far East Rand whose depositional subsidence could have been responsible for the folds of the Far East Rand.

(ii) *Convection current folds.*

It is possible that there are currents flowing horizontally below the crust of the earth. Such convection currents (Billings, 1942) would exert a drag on the overlying rock and set up a "couple".

The folds in the sediments of the Witwatersrand System on the Far East Rand are variable, some being asymmetrical and others symmetrical.

If convection currents with their resulting couple were the cause of the folds on the Far East Rand all the folds should be asymmetrical. This is not the case and it is most improbable that the folding was caused by convection currents if such currents do, in fact, exist.

(*c*) General Description of Folds and Probable Causes.

The folding on the Far East Rand may be classed as en echelon folding. The folds along the north-north-western flank of the synclinal basin pitch to the south-south-east and die out in about the centre of the synclinal basin, and the folds along the south-south-eastern flank pitch to the north-north-west and also fade out in approximately the centre of the synclinal basin. The folds from the north-north-western

flank do not join up with those from the south-south-eastern flank of the synclinal basin, but are offset and lie en echelon to each other.

Although there is not a well-defined line of depression there is a rather wavy line of depression roughly at 50° to the trend of the folding, that is the depression runs from east to west and is situated roughly in the centre of the Far East Rand basin.

The folding may be of the type that forms culminations and depressions, although there are no indications of culminations. It is possible that culminations were present but owing to erosion have disappeared.

The en echelon folding on the Far East Rand with the folds pitching to the south-south-east from the north-north-western flank and the folds from the south-south-eastern flank pitching to the north-north-west has naturally formed basins. There are two such perfectly formed synclinal basins, one near the western corner of Grootvlei (Pty.) Mines, Ltd. and the other in the south-western portion of Daggafontein Mines, Ltd. The long axes of these basins are parallel to the axes of the folding and trend from north-north-west to south-south-east (see Plate II). This type of folding is typical of that formed by a horizontal couple or by direct but not quite even pressure.

The major folds on the Far East Rand are not similar in form. On the northern flank of the synclinal basin the Van Dyk-Brakpan anticline is asymmetrical and the Modder "B"-Geduld Anticline symmetrical, that is the anticline on the west is asymmetrical. On the southern flank of the synclinal basin the Marievale-Vogelstruisbult Springs anticline is asymmetrical, and the Nigel-Sub Nigel Spaarwater-Vlakfontein and West Vlakfontein set of anticlines considered as a unit are symmetrical, that is the anticline on the east is asymmetrical.

Also in the northern half of the Far East Rand there is a large almost flat-bottomed syncline between the Van Dyk-Brakpan and the Modder "B"-Geduld anticlines, while in the southern half the syncline between the Marievale-Vogelstruisbult-Springs anticlines and the Nigel-Sub Nigel-Spaarwater-Vlakfontein-West Vlakfontein composite anticline is well defined with a definite axis to the trough. The major folds are considerably closer together in the southern half than those in the northern half of the Far East Rand.

This type of folding is certainly not the result of even pressure from the north-east, but it could be caused by uneven pressure from the north-east similar to folds formed by placing one's finger on a piece of damp blotting paper resting on a smooth surface and moving the fingers towards each other. As is the case in all experiments with damp clay or other materials the initial fold on compression is rather haphazard, but once this fold has been formed others follow. The folding on the Far East Rand, in which the folding is more pronounced on the western side in the northern portion of the area and more pronounced on the eastern side in the southern portion of het area, could be caused by uneven pressure from the north-east, the initial folds in the northern and southern portions of the Far East Rand being en echelon to each other.

The Witwatersrand System to the east of the Far East Rand is buried under a cover of the Karroo System and apart from a number of isolated boreholes very little is known of the area. These boreholes, to the immediate east of the Far East Rand, show, however, that the sediments of the Lower Witwatersrand System are very disturbed and most probably folded. It is not known how far the folding extends eastwards or what mass or masses were moved and caused forces that could be responsible for such uneven horizontal pressure. Folding of the Witwatersrand System similar to that of the Far East Rand is not present to the west of the Far East Rand. If the pressure that caused the folding was from the south-west it means that the whole

area underlain by the Witwatersrand System was moved to the north-east and that the north-eastern portion of this huge mass was folded. It is possible that this type of movement caused uneven pressure from the south-west and was responsible for the folds.

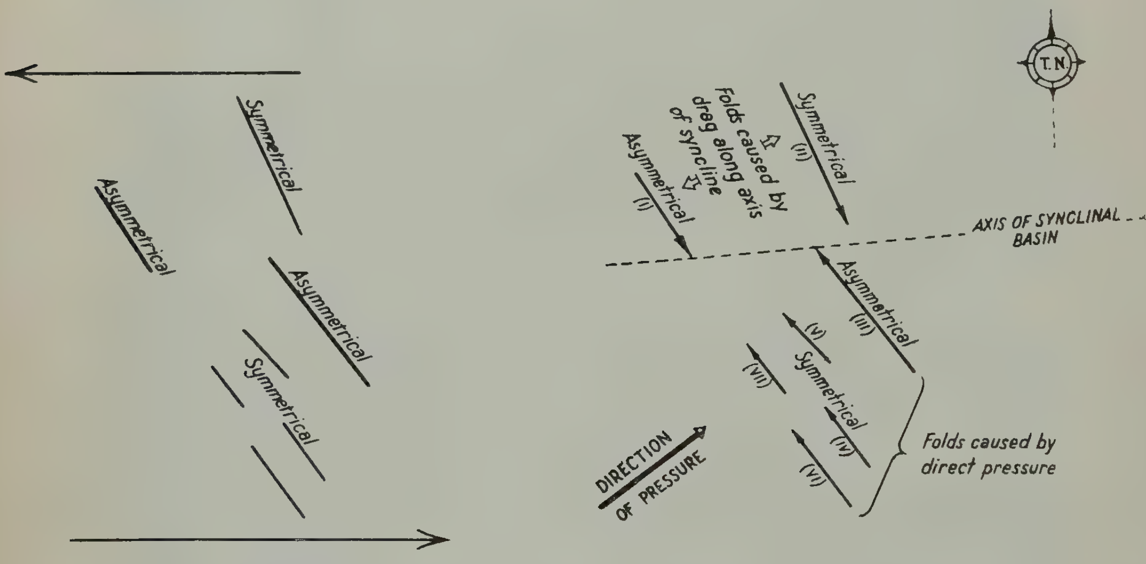
Although the folds of the Far East Rand could be caused by uneven pressure from the north-east or south-west there is another type of movement that will cause this en echelon folding, namely a horizontal couple.

If the northern half of the Far East Rand was moved to the west and the southern half to the east, or either of these blocks moved in these directions, folds roughly at 45° to the lines of movement would be produced.

If such a horizontal couple is the cause of these folds it is necessary to explain why the asymmetrical fold in the northern half of the Far East Rand is on the western side and the asymmetrical fold in the southern half on the eastern side; that is, the asymmetrical folds are furthest away from the active forces producing the couple as shown in Fig. 9 (*a*).

In the northern half the asymmetrical fold has been formed where the sediments are considerably thicker than those to the east where the asymmetrical fold has been formed, and in the southern half the asymmetrical fold has been formed where the sediments are thinner than those where the symmetrical folding is present to the west.

The variation in thickness of the sediments is thus not an explanation.



FAR EAST RAND
SKETCH SHOWING AXIS OF MAJOR FOLDS AT MAIN REEF HORIZON
(a)

FAR EAST RAND
SKETCH SHOWING AXIS OF MAJOR FOLDS AT MAIN REEF HORIZON AND POSSIBLE DIRECTION OF PRESSURE.
(b)

Fig. 9

It is possible that in the final foundering of the Witwatersrand tectonic basin a force was instituted which acted from the deepest to the shallowest portion of the tectonic basin, that is, the force acted in a north-easterly or easterly direction.

As in this case the tectonic basin was already formed, variations in the composition of the floor of the Witwatersrand System would play a large part in any distortion of a compressive force from the south-west or west.

The Nigel Old Granite was probably relatively stable during the tectonic basining; thus if the formations were compressed by a force from the west or south-west this mass of Old Granite would act as a buffer and distort the folding.

The outer edge of the Nigel Old Granite is folded, so that the mass did not act as a rigid buffer and portions at least were distorted.

This Old Granite, however, was most probably more stable than the formations into which it intruded, namely, the remainder of the Witwatersrand System floor of the area.

This Nigel Old Granite buffer would explain why the folds in the southern portion of the Far East Rand Basin are more closely folded than those in the northern portion, but does not explain the en echelon folds of the Far East Rand.

If the force from the west or south-west acted directly only on the formation to the south of the axis of the tectonic synclinal basin of the Far East Rand, namely, towards the Nigel Old Granite and the area to the south of this Old Granite, the folds of the southern half of the Far East Rand Basin would be caused by direct pressure against the Old Granite mass; this Old Granite buffer thus being the cause of the asymmetry of anticline (iii), Fig. 4 (*b*).

While the formations in the southern portion were being folded by such a force there would be a dragging effect on the formations in the northern portion of the Far East Rand Basin.

The effect of this drag would be that of a normal couple and would cause the formations in the northern portion of the basin to be folded (Lee, 1929).

The axis of the drag was probably determined by the axis of the tectonic synclinal basin of the Far East Rand which is approximately at 50° to the trend of the folds.

It is probable that the greatest amount of drag was on the west side of the northern flank of the Far East Rand syncline and was the cause of the asymmetry of the anticline. See Fig. 9 (*b*).

The force from the west or south-west may have been instituted as a result of the foundering of the Witwatersrand depositional basin or may have been caused by the relative upward movement of the Vredefort-Parys "ring", which as Brock suggests was not stable in pre-Transvaal System time.

Brock also suggests that the Sugarbush Fault was initiated in pre-Transvaal System time and was one of the faults responsible for the "Vredefort Ring". As will be indicated later in this discussion there is good evidence of pre-Transvaal post-Ventersdorp movement along this fault.

(*d*) General Descriptions of some Major Faults and Probable Causes.

Of all the faults on the Far East Rand the most common are those that strike roughly east and west. There are a large number of these faults and their strikes vary from 5° north of west and 5° south of east to 10° north of west and 10° south of east and average 6° north of west and 6° south of east (see Plate II).

Some of these faults are partly filled by igneous rock but more often there is no igneous rock present. Often in a fault that can be traced for a few miles various

portions are filled by igneous matter separated by portions that contain no igneous material.

The vertical displacement along the east to west trending faults varies considerably, as for instance in the fault that runs through the southern portion of Springs Mines, Ltd. and continues right through the northern portion of Vogelstruisbult G.M. Area, Ltd. In the western portion of Springs Mines, Ltd. the fault has an upthrow of about 100 feet on its southern side, further east there is no vertical displacement, and still further east the displacement is over 500 feet down on the south side of the fault. Near the common boundary of Springs Mines, Ltd. and Vogelstruisbult G.M. Areas, Ltd. the fault again has no vertical displacement, but about 4,000 feet east of this point there is a vertical displacement of 400 feet upwards on the south side of the fault. Further east there is again no displacement and again on the boundary between Vogelstruisbult G.M. Areas, Ltd. and East Daggafontein Mines, Ltd. there is a displacement of 300 feet downwards on the south side of the fault. The crest of the Marievale-Vogelstruisbult-Springs anticline (iii), Fig. 4 (*b*), has been displaced about 2,000 feet to the west on the north side of the fault.

As mentioned previously the gold content of the Main Reef is variable but the values run in definite "shoots" which trend from north-west to south-east. In the north-eastern portion of Vogelstruisbult G.M. Areas, Ltd. these shoots have been displaced 3,000 feet to the west on the north side of the fault. This same fault has been encountered in the workings on the May Reef and here too there is a large displacement; the reef on the north side of the fault being displaced a considerable distance to the west.

The principal movement on the east to west trending fault is horizontal. This large horizontal displacement is responsible for the variation in the vertical displacement along the strike of the fault, the movement having taken place after the major folds had been formed. The flanks of the anticlines on either side of the fault are not, however, exactly the same, which shows that folding continued although to a considerably less degree after the faulting.

There are numerous faults of this type over the whole of the Far East Rand, although in the northern portion of the area it is not possible to determine their horizontal displacement with any degree of accuracy as the reef shoots in these parts are not well defined. There are some, however, in which this movement can be seen by the displacement of the axes of the anticlines and also the axes of the synclines; as, for instance, the fault which runs through Grootvlei Pty. Mines Ltd., the southern portion of East Geduld Mines, Ltd., and the northern portion of New State Areas, Ltd. (see Plate II). The axes of the anticline (ii) and the syncline to its east have been displaced to the west on the north side of the fault.

In the southern portion of the Far East Rand where the reef "shoots" are more definite, especially on the Sub Nigel, Ltd. and Vlakfontein G.M. Company, Ltd., these displacements are clearly visible both on the coloured value contour maps and on the ordinary maps showing stoping.

On the Sub Nigel, Ltd. these east to west trending faults, besides showing horizontal movement, clearly show that the sediments between these parallel faults were folded still further after the faulting. This post-fault folding is on a small scale, the vertical difference between crest and trough seldom being more than 200 feet and usually considerably less. In many instances the minor puckers on one side of an east to west trending fault are very different from those on the other side of the fault.

Owing to this continuation of folding after faulting it is not possible to determine accurately the horizontal displacement along these east and west trending faults,

and this is especially the case where there is only a small amount of horizontal displacement. In one instance on the Sub Nigel the horizontal displacement on one of these faults is 100 feet to the west on the north side of the fault in one place and nil in another place about 3,000 feet away. Variations in horizontal displacement similar to the above are plentiful. These variations in horizontal displacement, which are determined by the displacements of the "shoots", are due to puckers being formed independently on either side of a fault.

All these faults with horizontal displacements, which trend roughly from east to west, are approximately at 45° to the axes of the folds, and would result from pressure from the north-east or from the south-west or would be the shears resulting from a couple acting parallel to the trend of these east and west trending faults.

(*e*) The Sugarbush Fault.

The Sugarbush Fault has been described by Rogers in his memoir on the Geological Map of the country around Heidelberg, and the following is a discussion on the relationship of this fault to the folds of the Witwatersrand and Ventersdorp Systems on the Far East Rand.

The fault is situated about seven miles south of Heidelberg and on surface is a rather wavy line, but west of Balfour and south of Heidelberg trends from a few degrees south of west to a few degrees north of east. On the farm Rietbult Estates 13, about 3 miles north-east of Balfour, the fault splits and the strike of both these faults is from west-north-west to east-south-east. The northern of these two faults is covered by Karroo sediments, but as can be seen from Rogers' map it must be present.

The Sugarbush Fault has a vertical displacement downwards on the south side, which varies from 3,500 feet on the farm Goedverwachting 306 in the west, to 15,000 feet on the farm Blinkpoort 208 in the east.

One remarkable feature of the fault in the Heidelberg-Balfour area is that on its north side both the sediments of the Witwatersrand System and lavas of the Ventersdorp System have been folded, and on its south side in this area these formations are not folded. Also on its northern side the regional dip of the Witwatersrand System is to the west or north-west and on its south side this formation dips almost due north.

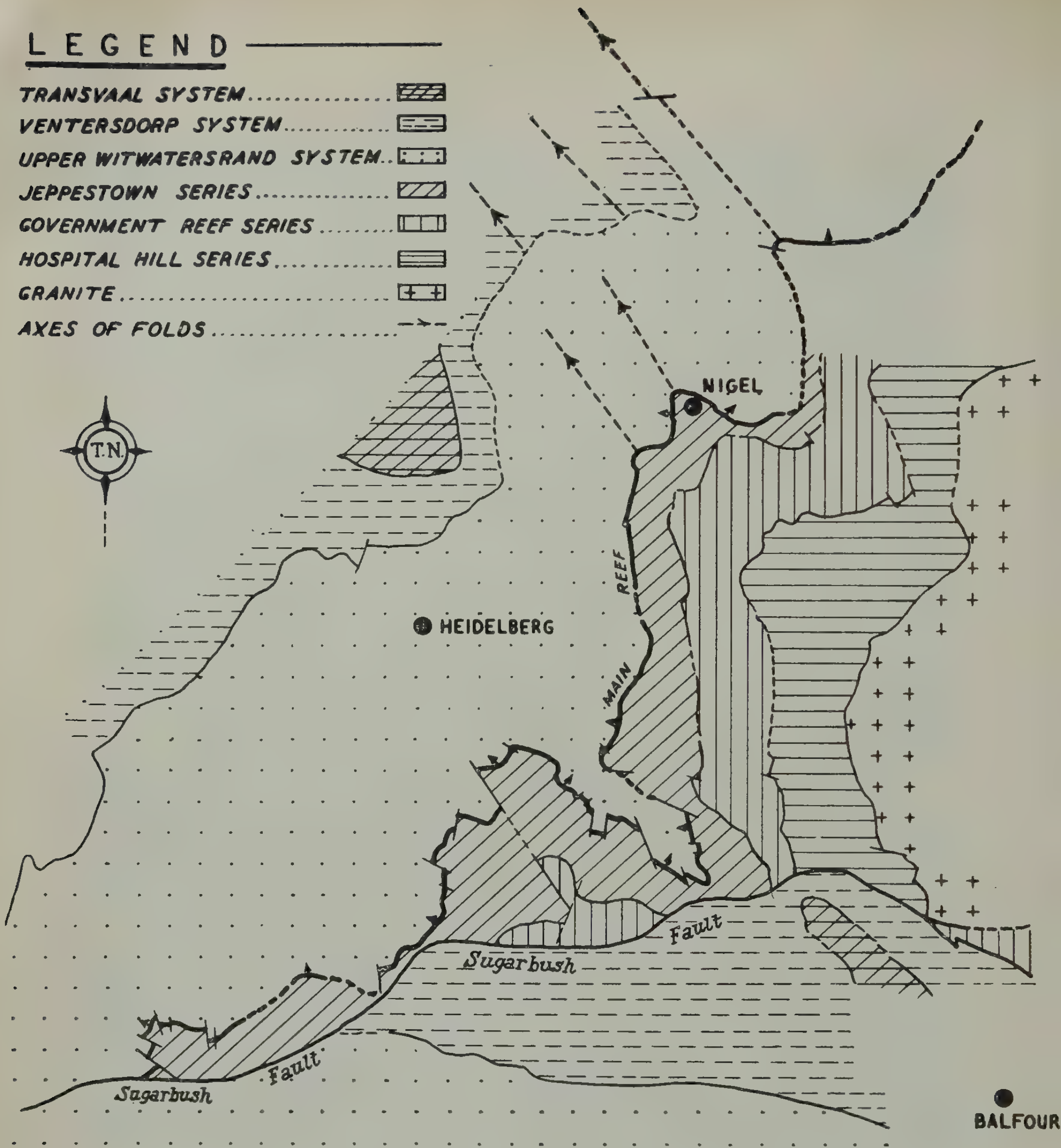
The dip of the fault is not known definitely, but in places appears to be to the south. Rogers states that the dip is probably almost vertical or steeply inclined to the south and that the displacement is that of a normal fault.

As Rogers states, the fault nowhere cuts the Transvaal System as a whole, but movement certainly took place after the deposition of the Black Reef and Dolomite Series of the Transvaal System and is most probably post-Transvaal System in age.

The folds on the north side of the Sugarbush Fault are a continuation of those encountered at the Main Reef horizon on the Far East Rand and are of pre-Transvaal System age. See Sections III and IV, Plate III. These folds terminate on the Sugarbush Fault. To the immediate south of the fault in this area the Witwatersrand System is not folded. This is not the result of normal faulting.

The only explanation of these contradictory facts is that movement on the Sugarbush Fault occurred in at least two different times and directions.

If during the forming of the folds of the Far East Rand by pressure from the west or south-west, pressure was not relieved sufficiently by this puckering, relief would have to have come by fracturing. If the Sugarbush Fault was a result of such relief of pressure and the folded portions of the Witwatersrand and Ventersdorp Systems on the south side of the fault were moved to the east, and relatively undisturbed formations moved into the position formerly occupied by the folded formations, it



MAP OF COUNTRY AROUND HEIDELBERG.
(AFTER A.W. ROGERS.)
SHEWING IN ADDITION AXES OF FOLDS AT MAIN REEF HORIZON.

Fig. 10

would explain why the sediments and lavas on the north side of the fault in the Heidelberg area are folded and comparable folds are not present on the south side of the fault in this vicinity.

The trend of the folds on the north side of the Sugarbush Fault is roughly at 45° to the regional strike of the Fault in this area.

To the north of Balfour, that is where the Sugarbush Fault abutts against the Nigel Old Granite, the strike of the fault changes to east-south-east.

The schists of the Primitive System on the farm Rietfontein 72 may be part of a fringe of schists on the edge of the original pluton of the Nigel granite similar to the schists around the Johannesburg-Pretoria Old Granite, and if this is the case (which unfortunately cannot be proved as this area is covered by Karroo beds) they probably represent a line of weakness and determined the strike of the Sugarbush Fault which on the farms Rietfontein 72 and Rietbult Estates 13 changes its direction to the east-south-east.

Whether the above is the explanation of the structures on either side of the Sugarbush Fault or not it is obvious that there must have been considerable horizontal movement along the Sugarbush Fault in pre-Transvaal System time. Du Toit (1939) has indicated that the northern edge of the Nigel Old Granite mass lies almost due east of the centre of the Far East Rand basin.

As horizontal displacements comparable in magnitude are not found in the Far East Rand, especially along the prolongation of lines from the northern end of the Nigel Old Granite parallel to the strike of the Sugarbush Fault, it would appear that the compressive forces that causes the folding were stronger to the south of the Far East Rand basin than to the north.

This type of pressure would cause couple action and it is possible that the folds of the Far East Rand are due to the drag along such a couple or couples.

After the horizontal movement on the Sugarbush Fault there was probably a certain amount of erosion and the Black Reef and Dolomite Series were then deposited on this eroded surface. Subsequent to their deposition there was vertical movement on the Sugarbush Fault and this is the movement that, as Rogers states, was post-Dolomite Series and probably post-Transvaal System.

There are indications all over the Far East Rand and Heidelberg and for that matter the whole of the area underlain by the Witwatersrand System that there were post-Transvaal System movements.

The vertical displacement on the Sugarbush Fault is probably related to these post-Transvaal System movements.

(*f*) Summary and Discussion.

The axis of the synclinal basin of the Witwatersrand System on the Far East Rand is roughly parallel to the strike of the Sugarbush Fault.

The numerous east to west trending faults are approximately parallel to the axis of the synclinal basin and the strike of the Sugarbush Fault.

The axes of the major folds of the Witwatersrand System on the Far East Rand are at approximately 45° to the east-west trending faults, to the axes of the synclinal basin, and to the strike of the Sugarbush Fault.

The axes of the minor folds are at roughly 45° to the strike of the east to west trending faults.

The major folds of the southern sector of the Far East Rand terminate on the Sugarbush Fault.

The minor folds on either side of the east to west trending faults often do not correspond and have been formed independently. There was movement to the east on the south side of these faults relative to the blocks on the north of the faults.

The major symmetrical and asymmetrical folds in the northern sector of the Far East Rand pitch to the south-east and do not correspond to the folds of the southern sector which pitch to the north-west. These folds indicate that they were formed independently on either side of the axis of the synclinal basin of the Far East Rand.

All the above facts indicate that the southern portion of the Far East Rand Basin moved to the east relative to the northern portion.

The minor folds and east to west trending faults are sympathetic to the major folds and faults, and as the relative movements along these minor faults are known, are further proof of the relative movement along the Sugarbush Fault and also the relative movements on either side of the axis of the synclinal basin.

It is very likely that the compressive force from the south or south-west was strongest in the vicinity of the Sugarbush Fault, resulting finally in the large horizontal displacement along this fault, and that the trough of the synclinal basin of the Far East Rand became the axis of a couple as the compressive force acted directly only on the southern portion of the Far East Rand Basin.

Pressure from the west or south-west and a resulting couple probably determined the trend and formation of the major folds.

During the major folding sympathetic faults occurred roughly at 45° to the trend of the folds and the horizontal movement along these faults set up couples which in many places caused minor puckers and folds.

Relative Ages of Basin and Folds.

On the Far East Rand the basining of the Witwatersrand Sediments and Ventersdorp Lavas was probably caused by subsidence immediately after or during the latter part of this extrusive phase of the Ventersdorp System.

The basining occurred definitely in pre-Transvaal System time.

The major folding of both the Witwatersrand and Ventersdorp Systems also took place in pre-Transvaal System time (see Sections I and IV).

The basin and major folds on the Far East Rand were thus instituted during the same geological age.

As has been shown it is most unlikely that the folds and the basin of the Far East Rand are due to the same causes, although forces resulting from the basining may have caused the folding.

On Grootvlei Mine, Ellis (1943) has shown that a major Ventersdorp dyke is displaced a considerable distance to the east on the south side of the large east to west trending tear fault, which fault, as has been shown, is related to the folding.

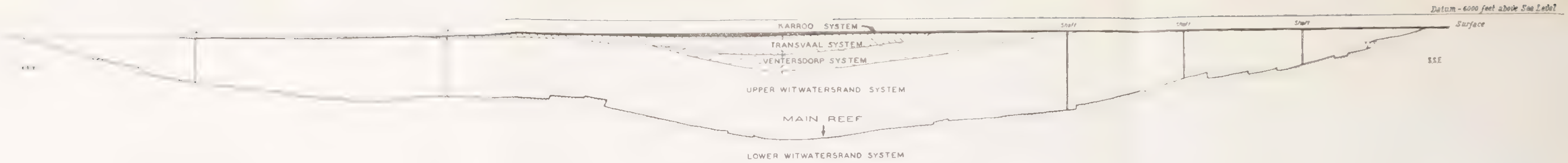
The age of this "Ventersdorp dyke" is not known and may have occurred before, during or after the extrusive phase of the Ventersdorp System.

If as is probably the case this dyke was contemporaneous with the extrusives of the Ventersdorp System, the fault that displaces the dyke would be post extrusives and thus probably post basining of the sediments and lavas.

The axis of the synclinal basin of the Far East Rand probably determined the en echelon folds of the area thus indicating that the basining of the sediments preceded the folding.

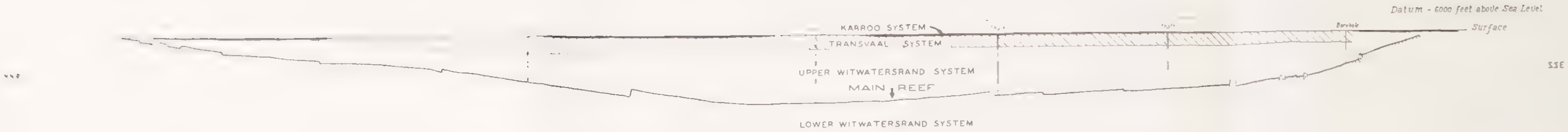

— FAR —
EAST RAND
MAP SHOWING MAIN REEF CONTOURS
SCALE 1 : 50 000
DATUM OR ZERO LINE IS 6000 FEET ABOVE MEAN SEA LEVEL
LEGEND
N
BRAKPAN MINES LTD
EAST DAGGAFONTEIN
MINES LTD.
WEST VLAKFONTEIN
G.M. CO. LTD
WEST SPAARWATER LTD

PLATE II



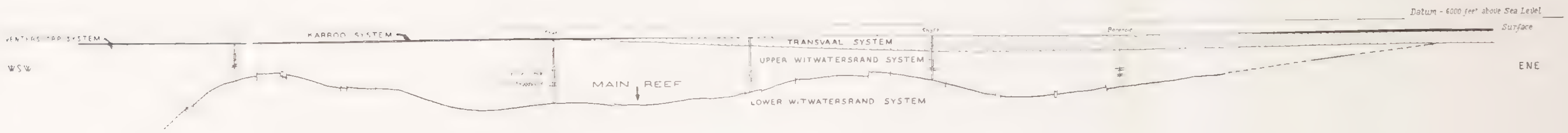
(SECTION I)

SECTION FROM NEW KLEINFONTEIN CO. LTD. TO NIGEL G.M. CO. LTD.
SHOWING MAIN REEF HORIZON.



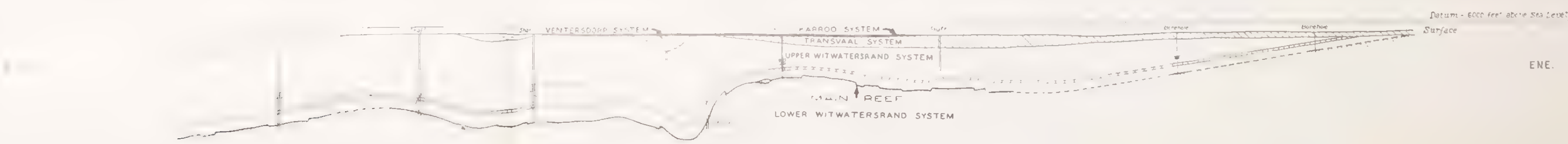
(SECTION II)

SECTION FROM MODDERFONTEIN EAST LTD. TO MARIEVALE CONS MINES LTD.
SHOWING MAIN REEF HORIZON



(SECTION III)

SECTION FROM VAN DYK CONSOLIDATED MINES LTD TO WELGEDACHT EXPLORATION CO. LTD.
SHOWING MAIN REEF HORIZON AND UNCONFORMABLE RELATION BETWEEN TRANSVAAL AND WITWATERSRAND SYSTEMS.

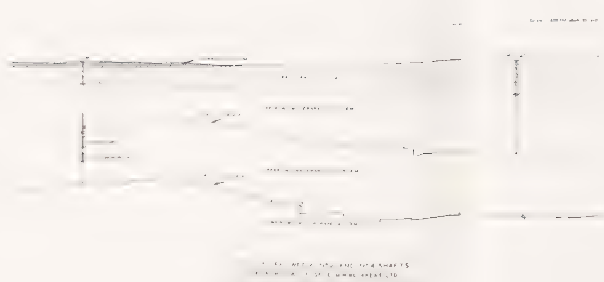


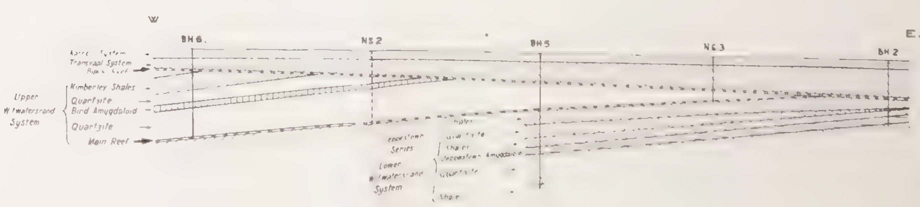
(SECTION IV)

SECTION FROM WEST VLAKFONTEIN G.M. CO. LTD. TO EAST DAGGAFONTEIN MINES LTD.
SHOWING MAIN REEF HORIZON AND RELATIONSHIP OF TRANSVAAL AND VENTERSDORP SYSTEM TO WITWATERSRAND SYSTEM.

1 0 1 2 3
Scale in Miles

(SECTION VI)





SECTION THROUGH HOLFONTEIN Nº 1
AND MODDERFONTEIN Nº 46.

1 0.5 0 1
Scale in Miles

PLATE III

The folds probably took some considerable time to form and it is very doubtful if there was any time-lag between the pouring out of the lavas and the basining of the sediments and lavas, at any rate sufficient time for the folds to have been formed.

The folding obviously took place later than the basining of the Witwatersrand and Ventersdorp Systems on the Far East Rand.

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MORPHOLOGICAL RECONSTRUCTION OF THE KIMBERLEY-ELSBURG SERIES, WITH SPECIAL REFERENCE TO THE KIMBERLEY GROUP OF SEDIMENTS IN THE EAST RAND BASIN

By

F. S. J. DE JAGER, M.Sc.

(Submitted July, 1952)

ABSTRACT

The ideal geological column of the Kimberley-Elsburg Series in the East Rand Basin is described, with particular reference to the Kimberley group of sediments.

A system of nomenclature has been devised, and it is suggested that it could also be used in other parts of the large structural basin, stretching from Johannesburg in the north to near Theunissen in the south, and from Klerksdorp in the west to Greylingstad in the east.

The stratigraphy of individual areas in the East Rand Basin is described in detail, and it is shown that certain stratigraphic units have remarkable regularity, maintaining their lithological characteristics over large areas, and persisting also into the Greylingstad-Balfour district, the Central Rand, the West and Far West Rand, the Klerksdorp area, and into the Orange Free State gold field.

In the East Rand Basin the May Reef is the principal gold carrier, and is economically important in certain mines. In the Orange Free State gold field the lowermost Kimberley reef is also of economic importance.

Three regional unconformities have been recognised in the part of the column extending from below the Kimberley Shales to above the May Reef. The May Reef covers the upper one, and owes its existence to this period of erosion. The history of this reef could be traced back to its parent rocks, in this case, stratigraphically older auriferous gravels. The author believes that the unconformity below the May Reef developed as a result of sub-aqueous erosion. The oldest erosion surface probably developed in the same way. The middle one developed largely on the land, but was subsequently submerged.

It is concluded that the sediments of the Kimberley-Elsburg Series were deposited in the marine neritic environment, i.e. in a sea of substantial but not excessive depth.

CONTENTS

I INTRODUCTION

A. General

Location of the Area

The East Rand Basin is part of the Witwatersrand gold field, the centre of the area lying near Dunnottar, some 25 miles to the east of the city of Johannesburg. The area studied in detail (sub-surface geology) measures approximately 22 miles along a north-south line, and about half that distance along an east-west line. The town of Heidelberg is situated near the southern edge of the Basin (Fig. 1).

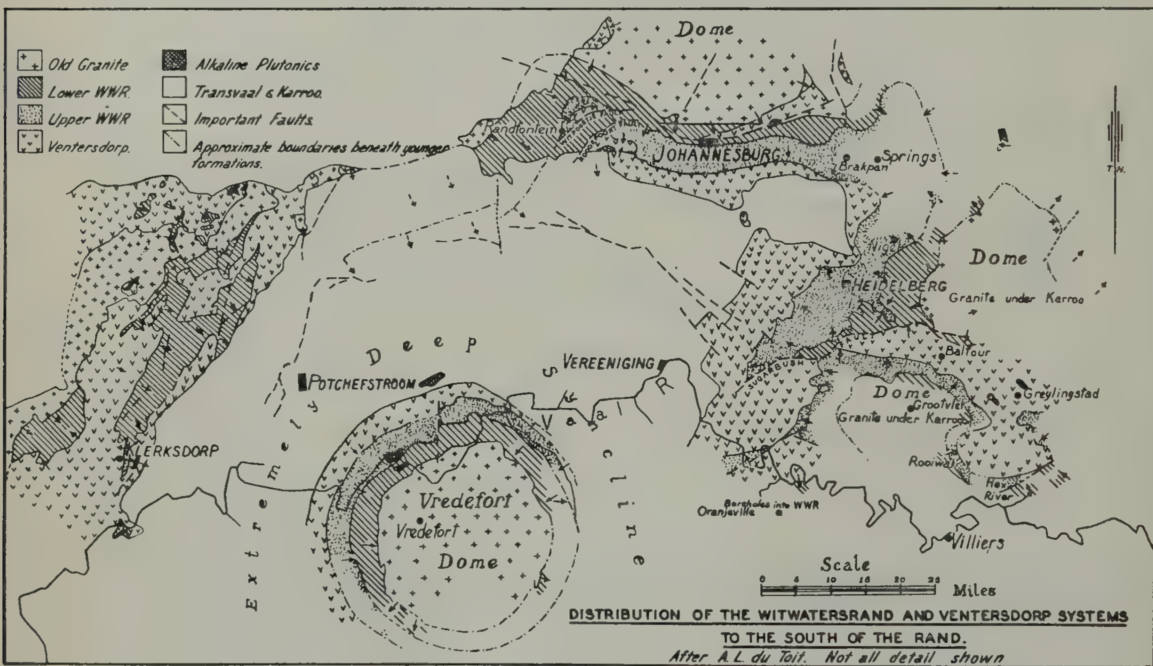


Figure 1

The Regional Basin

The Witwatersrand System with its auriferous deposits in the upper division has now been traced for a distance of some 200 miles to the south of Johannesburg to near Theunissen in the Orange Free State. In the Orange Free State gold field the rocks of the Witwatersrand System are nearly everywhere covered by younger formations.

Along an east-west line this large structural basin is defined by intermittent outcrops from around Klerksdorp in the west to Greylingstad in the east, a distance of about 150 miles.

Physical Features of the Witwatersrand

The general elevation of the Witwatersrand area is considerable, very little of it falling below 5,000 feet above sea level. Some of the highest points lie within the city of Johannesburg itself and reach nearly 6,000 feet. The Witwatersrand is part of the High Veld of the Southern Transvaal, and forms part of one of the main watersheds of the sub-continent. The northward flowing streams drain ultimately to the Indian Ocean, and the southward flowing streams find their way to the Atlantic. (Mellor, 1917 and Rogers, 1922.)

Age of the System

"The System is not fossiliferous, and is in all probability of Pre-Cambrian age"—Reinecke (1930).

B. The Thesis

The investigation was started in April, 1944, with a view to assess the economic possibilities of the Kimberley group of conglomerates.

Some of the gold mines on the Rand have been exploiting for many years the principal auriferous reefs at and near the junction of the upper and lower divisions of the System, and are now approaching old age. Some have been forced to close down. During recent years more and more attention has, therefore, been directed to the so-called "lesser" reefs, notably the Kimberley Reefs and Bird Reefs. A fair amount of success has been achieved in some areas.

As a result of the drilling programmes carried out during the past eight years in the East Rand area, largely under the supervision of New Consolidated Gold Fields, Limited, and as a result of the extensive underground development carried out by various companies, a great deal of information has been accumulated. The author also had the opportunity of describing the boreholes drilled on the Marievale and Nigel Mines.

For purposes of description and argument, and for reasons that are mainly lithological and economic, the East Rand Basin is divided into two sectors, a western and an eastern (Plate VII). The dividing line runs northward through the middle of the Nigel Mine, through a point not far east of Vogelstruisbult No. 3 Shaft, thence through Springs Mines in a north-westerly direction, and then through a point somewhere east of Government Gold Mining Areas. The geological history of the Kimberley-Elsburg Series in the western sector is, however, fundamentally the same as that in the eastern sector.

While the East Rand Basin yielded sufficient information for a comprehensive treatise, it was necessary to look further afield for the solution of various aspects of the problem. The investigation was therefore also carried into other areas: the Greylingstad-Balfour district lying to the south-east of Heidelberg, the Central Rand, the West Rand (Randfontein-Krugersdorp), the Far West Rand (i.e. the area around the Blyvooruitzicht Mine), the Klerksdorp area, and the area immediately to the south and south-east of Odendaalsrus in the Orange Free State. For evidence from the West Rand and Klerksdorp areas the author had to depend entirely on the literature.

The results of this regional investigation have exceeded expectations. It can now be stated with confidence that a regional correlation of individual stratigraphic units within the Kimberley-Elsburg Series exists.

In this endeavour to unravel the depositional history of the Kimberley-Elsburg Series on a regional basis, the results of the investigation in the East Rand Basin are used as a key to the solution of the problem over the entire structural basin. In this connection the pioneering work of Sharpe (1942-1945) on Government Gold Mining Areas should be mentioned. In the pages that follow, much attention is therefore paid to a great amount of detailed observation and deduction in the East Rand Basin.

Twenty-one years ago Reinecke (1930) stated that a study of the origin of the sediments of the Witwatersrand System appeared to be the most useful line along which an investigation into the mode of distribution of the gold in the reefs could proceed. The succeeding years have added force to his opinion, which to-day is universally accepted by stratigraphers and sedimentationists on the Rand.

II HISTORICAL REVIEW

A. Mining and Prospecting of the Kimberley Reefs

(*a*) *Greylingstad-Balfour District:*

This area is separated from the East Rand Basin by the Sugarbush Fault, which has a considerable downthrow to the south (Fig. 1).

In 1936 G. Carleton Jones correlated what appeared at the time to be the only surviving member of the Kimberley conglomerate group in the area south-east of Heidelberg with the highly payable intersection of Kimberley reef in the Daggafontein No. 3 Shaft.

Kimberley reef has been mined south of Greylingstad and Balfour since before the Boer War, and following the abandonment of the Gold Standard by the Union in 1932, considerable activity again took place in this district. Several small mining companies, notably the East Nigel Gold Areas, Limited, East Wits Gold Mining Areas, Limited, New Rand Reefs, Limited, and South-East Wits Gold Mining Company, Limited, carried out extensive development between 1933 and 1936 on what subsequently proved to be the May Reef. A portion of the former New Rand Reefs and South-East Wits was worked for a short period a few years ago under the name of the Hex River Gold Areas, Limited.

Westwards from the Greylingstad-Balfour sector the Kimberley reefs increase in number and robustness, forming conspicuous outcrops in the Malanskraal-Tweefontein area. Prospect trenches and pits are particularly numerous on Malanskraal, but an economic horizon was apparently not discovered.

(*b*) *The West Rand:*

This is the Randfontein-Krugersdorp area lying immediately to the north-west of the Witpoortje Fault.

In 1936 G. Carleton Jones briefly described the Kimberley reefs in this part of the Witwatersrand gold field: Luipaardsvlei Estates was exploiting the Battery Reef, West Rand Consolidated Mines the Pay Band or Battery Pay Band, and Randfontein Estates the Horsham Reef and Lindum Reef. Jones considered the Battery Reef, Battery Pay Band and the Horsham Reef to be one and the same horizon.

(*c*) *The East Rand Basin:*

1. *Outcrop Areas.*

The Kimberley reefs make conspicuous outcrops between Nigel and Heidelberg, and have been prospected extensively in the past, the more robust conglomerates having received most attention.

Nigel Mine has done some prospecting in the outcrop areas during the past few years.

A number of closed-in pits were discovered a few years ago near the May Reef outcrop in the southern portion of the Vogelstruisbult Mine, close to the common boundary with Marievale. On being re-opened some of them were actually found to intersect the May Reef some 30 to 50 feet below the surface, and at the bottom of one of the pits a winze followed the reef for a short distance. It appears that this prospecting must have been done many years ago, when the search for the extension of the Main Reef Group was in progress. Dating from about this same period is an old incline shaft in the Draaikraal sector of the Marievale Mine (Plate VII).

2. *Mines of the Anglo-American Corporation.*

The East Rand mines of the Anglo-American Corporation (Brakpan in the west to East Daggafontein in the east) started active prospecting of the Kimberley reefs about 1936, and between the years 1937 and 1943 a total distance of some 43,000 feet of boreholes was drilled from surface in this area (Jeppe, 1946).

A considerable amount of Kimberley reef development has been done on both Daggafontein and East Daggafontein, which has resulted in the delimitation of important pay shoots on the May Reef. Springs Mines achieved a little success in the north-eastern corner of the property.

3. *Government Gold Mining Areas* (*Johannesburg Consolidated Investment Company*).

During the period 1937 to 1943 Government Gold Mining Areas drilled a great many boreholes in exploration of the Kimberley group of conglomerates. They disclosed some very interesting values on the May Reef, and sporadic payable values on a few other bands.

4. *Vogelstruisbult Mine* (*New Consolidated Gold Fields*).

Following on the good results on Daggafontein and East Daggafontein Mines, Vogelstruisbult started underground development from their No. 1 Shaft early in 1940, and after an initial period of poor results in the immediate vicinity of the shaft, decided to advance a haulage eastwards, which resulted in the delimitation of important pay shoots.

Eleven boreholes were subsequently drilled in the outcrop area near the common boundary with the Marievale Mine, and a haulage advanced eastwards from No. 2 Shaft.

5. *Nigel and Marievale Mines* (the latter belonging to the Union Corporation Group) followed soon, initially by a period of drilling, and subsequently by underground development.

6. *The Western East Rand.*

During the years 1944 to 1949 a comprehensive drilling programme was carried out in the area from Vlakfontein Mine in the north to Witwatersrand Nigel

near Heidelberg in the south, under the supervision of New Consolidated Gold Fields.

A small amount of underground development has been done on Sub-Nigel, Vlakfontein and West Vlakfontein.

(*d*) *The Far West Rand:*

The May Reef and its characteristic footwall horizons have been recognised in borehole E.9.H near Blyvooruitzicht Mine on the Far West Rand.

B. Literature.

(*a*) Rogers (1922) briefly describes the geology of the Kimberley-Elsburg Series in the East Rand area, based mainly on field observations.

(*b*) du Toit (1939) remarks on the geology of the Kimberley-Elsburg Series.

(*c*) In 1943 Sharpe published a paper based on the results of the extensive drilling programme carried out on Government Gold Mining Areas, in which he suggested sub-divisions and a system of nomenclature, which subsequently proved of invaluable service to many of the East Rand mines.

(*d*) In 1949 Sharpe published another paper, suggesting a mode of deposition for the rocks of the Upper Witwatersrand System, based partly on his observations of the Kimberley-Elsburg Series on Government Gold Mining Areas. Replying to a discussion on his paper by the present author (1949), Sharpe produced more interesting information, which cast light on the origin of the rocks of this System.

(*e*) In 1950 Whiteside published a paper on the geology of the Kimberley-Elsburg Series in the East Rand mines of the Anglo-American Corporation.

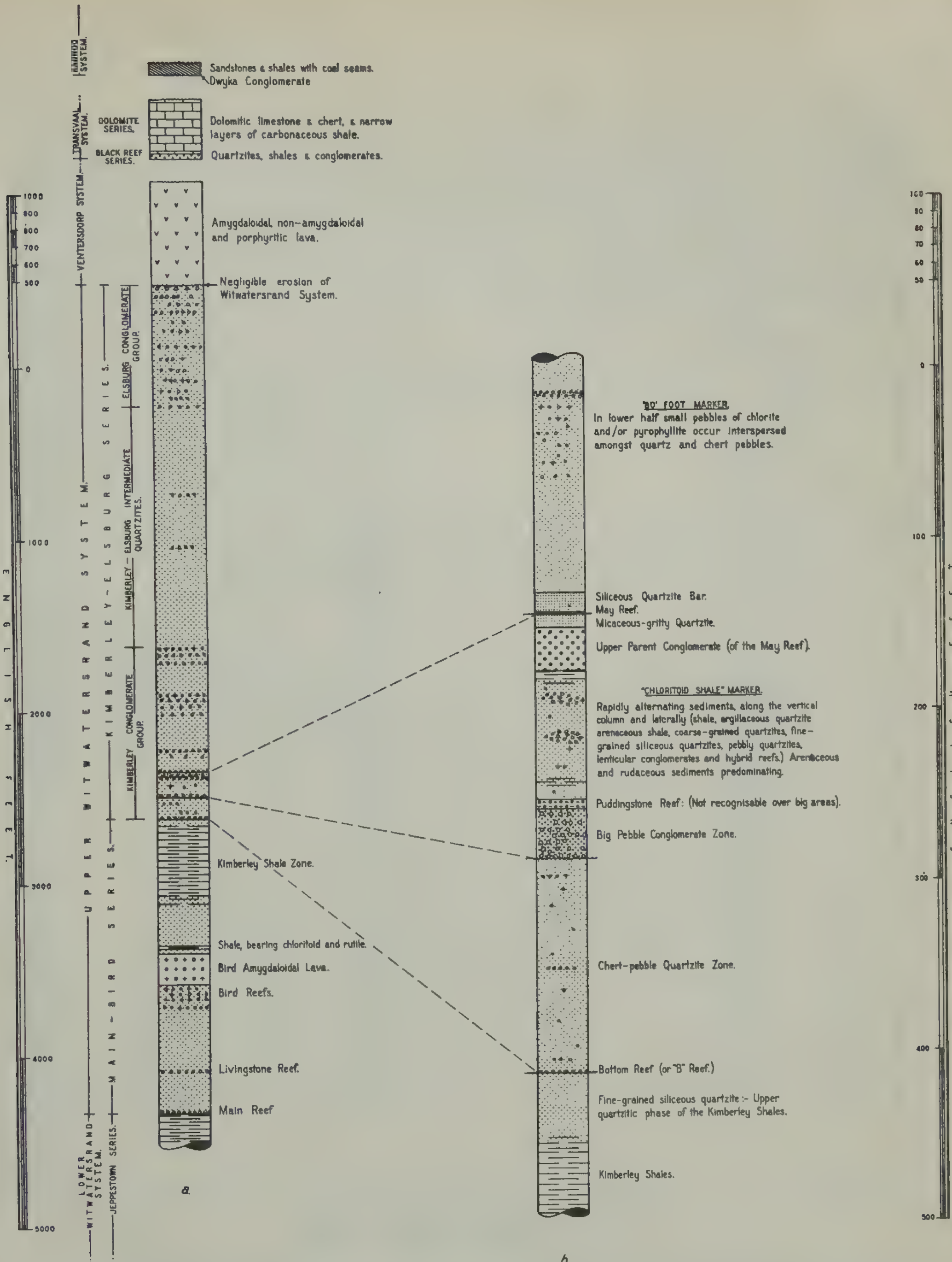
(*f*) Recent publications on the geology of the Upper Witwatersrand System in the Orange Free State and Klerksdorp areas refer to the Kimberley-Elsburg Series.

III BRIEF REVIEW OF THE WITWATERSRAND SYSTEM IN THE EAST RAND BASIN, WITH SPECIAL REFERENCE TO THE UPPER DIVISION.

Broadly the System is divided into an upper and a lower division, based partly on the difference in lithology, and partly on the fact that a period of erosion preceded the deposition of the only surviving member of the Main Reef Group. This reef lies at the junction of the competent quartzites of the upper division, and the incompetent shales of the lower division (Plates I and II).

The upper division, which in the areas of maximum development reaches a thickness of a little over 5,000 feet in the East Rand Basin, includes one persistent shale zone up to 500 feet thick, and a less persistent chloritoid-bearing shale zone a little higher up in the geological column. It includes also a zone of amygdaloidal lava, varying in positional distance from a few feet to several hundred feet below the Kimberley Shales.

The upper division is divided into the Main-Bird Series and the Kimberley-Elsburg Series, the Elsburg group of conglomerates occurring at the top of the System. In the past, the upper limit of the Main-Bird Series was generally placed at the base of the Kimberley Shales, but because the passage upward into the Kimberley Shales is a gradual one, and because it is considered that the lowermost Kimberley reef lies unconformably on the Kimberley Shale Zone, it appears logical to group the latter with the Main-Bird Series.



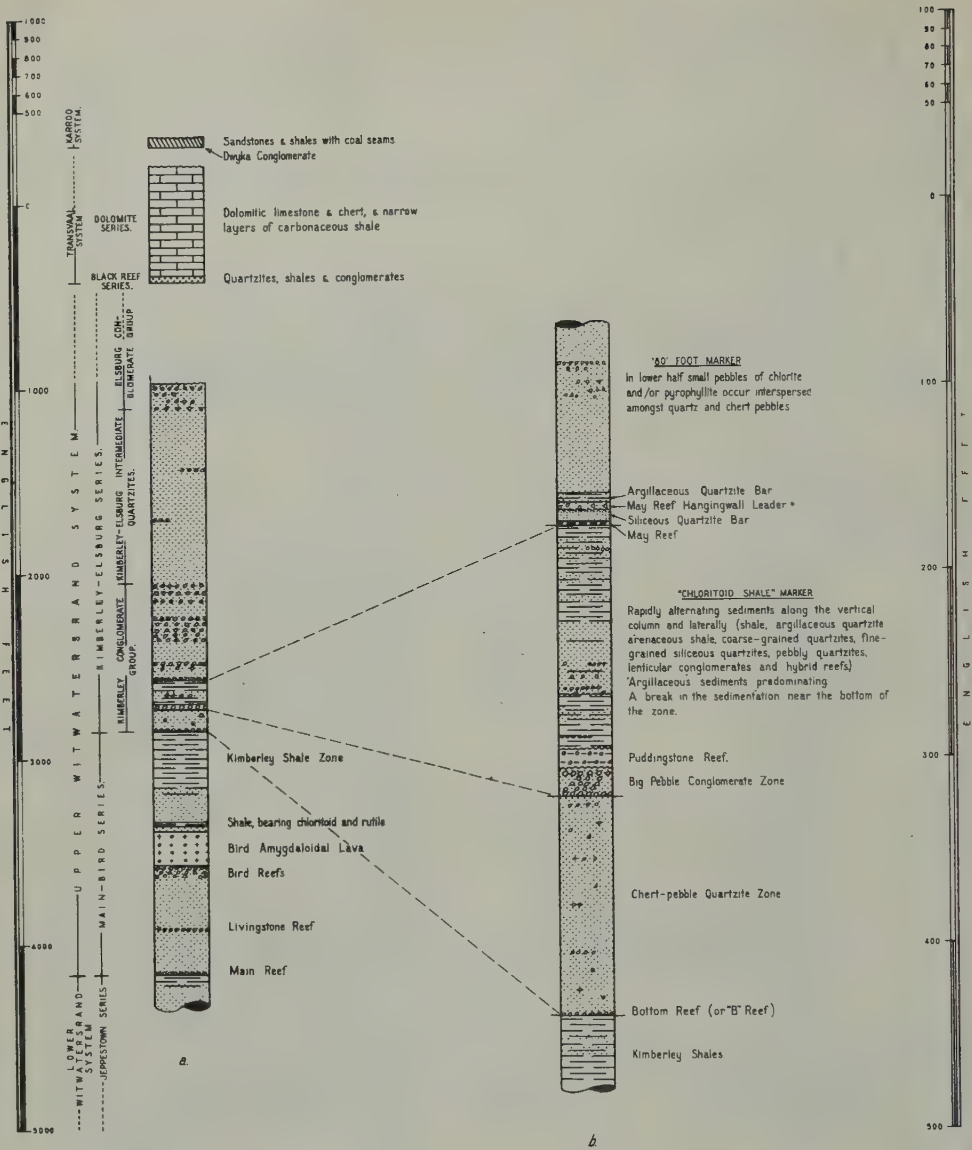
THE WESTERN EAST RAND.

a). Ideal Section of the Upper Witwatersrand System:-
(The total distance between the base of the Ventersdorp Lava and the Main Reef Leader is often increased by 400 to 500 feet as the result of a sill-like intrusion either in the Kimberley Shales or in the Main-Bird Quartzites.)

b). Ideal Section of the Kimberley Economic Zone:-
Top of '80' Foot Marker down to and including the Bottom Reef.

Note:- 1) There exists a perfect gradation between the Kimberley Shales and the Main-Bird Quartzite. The Kimberley Shales are therefore bracketed with the Main-Bird Series.
2) Section b).:- Interruptions in sedimentation within the Kimberley group of sediments, depicted thus :-

PLATE I



THE EASTERN EAST RAND

a) Ideal Section of the Upper Witwatersrand System:–
(The total distance between the May Reef and the Main Reef Leader is often increased by 150 feet to 200 feet as the result of a sill-like intrusion either immediately above or immediately below the Bird Amygdaloid.)

b) Ideal Section of the Kimberley Economic Zone:–
Top of '80 Foot Marker down to and including the Bottom Reef.

Note:– 1) There exists a perfect gradation between the Kimberley Shales and the Main-Bird Quartzite. The Kimberley Shales are therefore bracketed with the Main-Bird Series.
2) Section b:– Interruptions in sedimentation within the Kimberley group of sediments, depicted thus:– ~~~~

The Kimberley-Elsburg Series is therefore initiated by coarse sediments, including one inconstant shale zone (often chloritoid-bearing) as described above. The Series is on the average 3,100 feet thick in the Western East Rand, but in the eastern sector no full measurement is available owing to the periods of erosion that preceded the deposition of the rocks of the Transvaal and Karroo Systems (Plates I and II).

The Kimberley conglomerate group occurs in the lower third of the Series and has an average thickness of 1,000 feet in the western sector and 800 in the eastern sector.

The Elsburg conglomerate group has an average thickness of 700 feet in the East Rand Basin.

The System has been metamorphosed regionally, an interesting feature being the occurrence of ubiquitous authigenic rutile in the form of minute crystals, frequently twinned, and crystal aggregates, generally in greater abundance in the more argillaceous phases.

The Kimberley Economic Zone.

The Kimberley Economic Zone is defined as the series of sediments extending from the top of the "80" Foot Marker to the Bottom Reef on the Kimberley Shale Zone. The May Reef is the principal gold carrier in this series, although the top band of the "80" Foot Marker gave sporadic payable values on Government Gold Mining Areas, while the Bottom Reef is of economic importance in parts of the Orange Free State gold field.

The Kimberley Economic Zone embraces the lower half of the Kimberley conglomerate group (Plates I and II). The different stratigraphic units display a remarkable regularity, maintaining their lithological characteristics over large areas. The thicknesses of the individual units are, however, often extremely variable. Most of the conglomerate horizons and groups have been recognised both in the western and eastern sectors. Exceptions to these general rules are:

1. One of the most constant footwall conglomerates of the May Reef in the Western East Rand disappears eastwards, and has thus far not been encountered in the eastern sector.
2. The other exception concerns a suite of sediments which displays a very marked change of facies from predominantly coarse to predominantly fine sediments, when traced from west to east across the Basin. This is the suite that includes the chloritoid shales.

The recognition of the individual stratigraphic units depends largely on a knowledge of the pebble types, their sizes and shapes, and their distribution along the vertical column.

The May Reef is the economic horizon in the Kimberley conglomerate group in the East Rand Basin. Over comparatively large areas in the eastern sector, and in portions of Government Gold Mining Areas in the western sector its gold content is sufficiently high to warrant mining.

IV NOMENCLATURE AND CORRELATION OF INDIVIDUAL STRATIGRAPHIC UNITS.

In his two very interesting papers published in 1943 and 1949, Sharpe suggested a symbol classification for the Kimberley beds on Government Gold Mining Areas, where the conglomerate zones are very strongly developed.

In 1943 he recognised upper, middle and lower divisions:

(*a*) The lower division includes the Kimberley Shales and associated quartzitic phases.

(*b*) The middle division includes in ascending sequence:

1. A thick zone of pebbly quartzites in which angular chert pebbles predominate—the M.K.3 quartzites. The letters M.K. designate "Middle Kimberley".
2. A thick and robust conglomerate zone containing erratic payable gold values. This is his M.K.2 conglomerate zone.
3. His channel deposits: a heterogeneous assemblage of shales, quartzites and conglomerates.

(*c*) In his upper division he recognises cycles or periods of sedimentation, giving rise to groups of conglomerates and groups of quartzites. There are 9 such periods, U.K.9 to U.K.1, the last named occurring at the top of the group. The abbreviation U.K. designates "Upper Kimberley". The conglomerate zones correspond to the odd numbers. It apears, however, that his U.K.1 conglomerate zone, which is only sporadically developed on Government Gold Mining Areas and not found away from that mine, could be more logically grouped with the Kimberley-Elsburg Intermediate Quartzites. That would reduce the thickness of the Kimberley conglomerate group on this mine to about 1,200 feet, which compares better with the average thickness of the group elsewhere in the Western East Rand.

The U.K.9 includes the two principal gold-bearing horizons, occurring respectively at the top and bottom of the 50-foot-thick conglomerate zone. The upper one, U.K.9A, is economically the more important. Sharpe considers the lower horizon, U.K.9C, to be the equivalent of the May Reef of Daggafontein and East Daggafontein Mines. This is also the point where, in 1949, he placed his single "hiatus" in the Kimberley group of sediments.

He subsequently agreed with the author (1949) that the U.K.9A Reef corresponds to the May Reef of the eastern sector.

In his second paper (1949) he groups the Kimberley Shales and the coarse sediments above (up to, but excluding the May Reef) with the Main-Bird Series. Baines (1949) follows the same general pattern of correlation, and suggests the name "Kimberley Stage of the Main-Bird Series" for the Kimberley Shales and associated quartzitic phases. According to this grouping the May Reef rests unconformably on the Main-Bird Series, while the sediments above, up to the base of the Ventersdorp Lava, are grouped under the name Elsburg Series.

In his contribution to Sharpe's second paper the present author (1949) questions Sharpe's interpretation of a certain suite of sediments as typical channel deposits, and discusses the unconformable relationship between the May Reef and the older gravels. It was realized then that the history of the formation of the May Reef could be traced back to its parent rocks, in this case, stratigraphically older gravels. Instead of Sharpe's symbols a system of descriptive names was suggested. But it must now be admitted that some of the names advanced were not very suitable, and that one at least was a complete misnomer.

Whiteside (1950) describes the subsurface geology of the Kimberley-Elsburg Series in the Brakpan-Springs area. He adheres very closely to Sharpe's original classification and nomenclature. He finds it difficult to define the May Reef in the western sector as compared to the easily recognisable gold carrier of the eastern sector. This has been a stumbling-block to all investigators, but the detailed stratigraphic

study which was possible on the Nigel and Sub-Nigel Mines, in the author's opinion, clarified the position.

Whiteside gives a very interesting section through the May Shaft on the boundary between Daggafontein and East Daggafontein Mines. He argues very convincingly for a deep erosion channel below the May Reef in this area, filled with a heterogeneous assemblage of fine and coarse sediments in the lower half, and with fine-grained argillaceous quartzites and shales in the upper half.

As opposed to Sharpe, he groups the Kimberley Shales with the Kimberley beds.

It is thus evident that for those unacquainted with the problem the question of correlation and nomenclature has become very confusing. A modified system of descriptive nomenclature is thus proposed here, and it is suggested that it could be used over the entire structural basin, possibly in a modified form in some areas. This system of classification and nomenclature is fully set out on the accompanying table of correlation. The Kimberley Shales and associated quartzitic phases are bracketed with the Main-Bird Series, and the Kimberley conglomerate group divided into upper, middle and lower divisions, which are based on the three regional unconformities that have been identified.

Latterly, repeated reference has been made in the literature to interruptions in the sedimentation within the upper division of the Witwatersrand System. Sharpe recognises one such break in the Kimberley beds on Government Gold Mining Areas (below his U.K.9C Reef), but it now appears to be a very local one.

Twenhofel (1936) defines an unconformity as "the surface separating two distinct deposits, this surface representing a time during which no deposition took place over the area where the unconformity exists, or, if deposition did take place, the deposits were removed before the deposition of the strata above the unconformity", and again "an unconformity represents a rate of deposition that is zero, or if something is removed, a rate that is less than zero".

As applied to the Kimberley-Elsburg Series, the term unconformity is here used for a regional surface which represents a rate of deposition that is zero in some areas, and less than zero in adjacent areas.

Within the Kimberley group of sediments there exist, apart from minor local interruptions owing to scour, three regional unconformities:

1. the Bottom Reef or "B" Reef covers the lower one;
2. the Big Pebble Conglomerate was deposited on the next higher one, While
3. the May Reef blankets the upper one.

In some areas we have the cumulative effect of these three surfaces of erosion, when the May Reef comes to rest directly on the Kimberley Shales.

V THE KIMBERLEY-ELSBURG SERIES. IDEAL GEOLOGICAL COLUMNS IN THE EAST RAND BASIN.

The complete geological column of the Kimberley-Elsburg Series as constructured from a large number of borehole and shaft sections and observations underground is described here. It has been thought best to consider the eastern and western sectors of the Basin together in this chapter, and to produce, in effect, one column, although there are various differences in the geology of the two sectors. These differences are pointed out in the relevant paragraphs. The ideal geological columns in the two sectors are shown on Plates I and II.

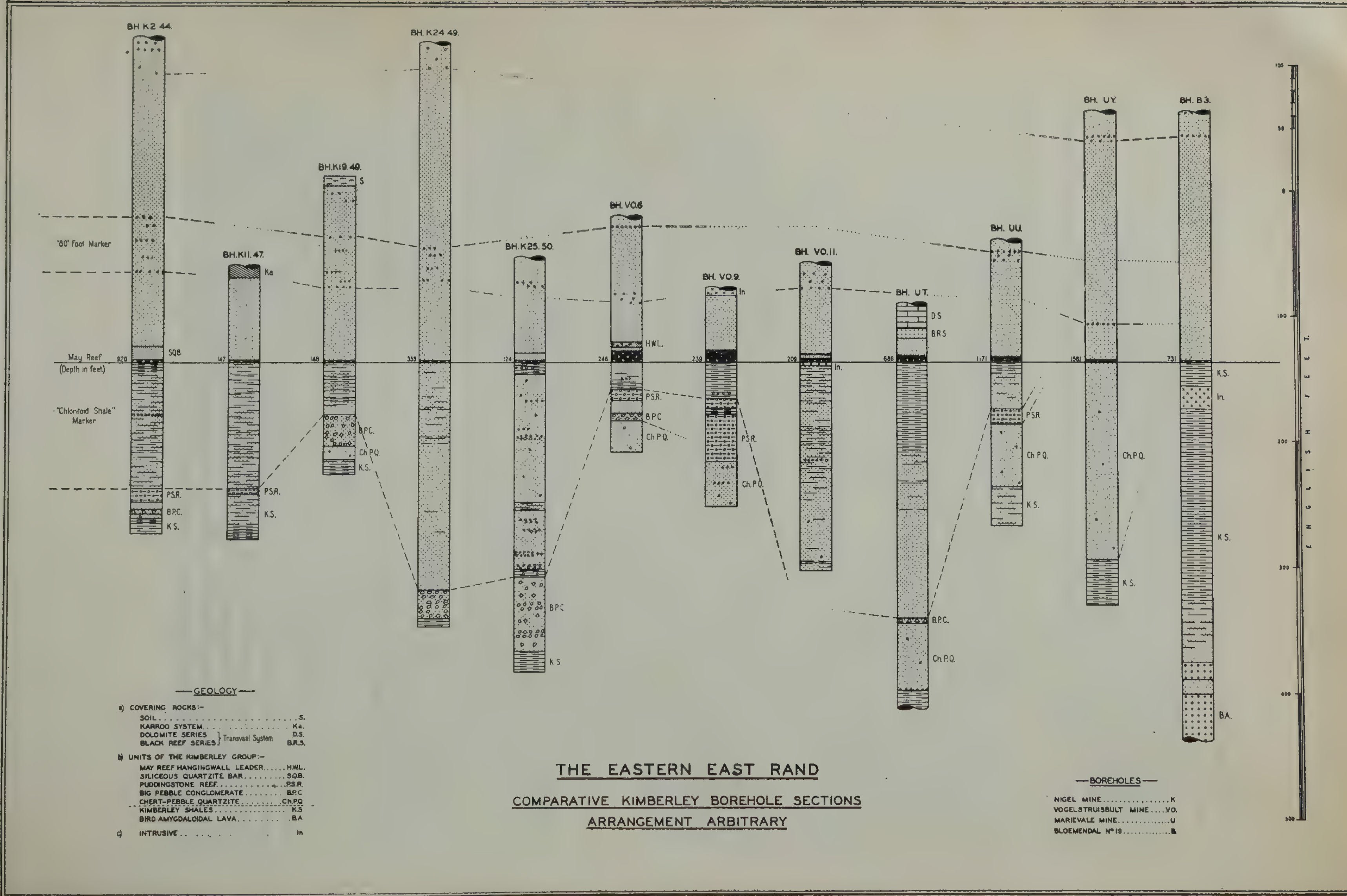

THE EASTERN EAST RAND
COMPARATIVE KIMBERLEY BOREHOLE SECTIONS
ARRANGEMENT ARBITRARY
BH K2 44.
BH.KII.47.
BH.KI9.49.
BH.K24 49.
BH.K25.50.
BH.VO.6
BH. VO.9.
BH. VO.II.
BH. UT.
BH. UU.
BH. UY.
BH. B3.
"80' Foot Marker
May Reef (Depth in feet)
"Chloritoid Shale" Marker
920
147
148
355
124
248
239
209
686
1171
1581
731
SQB
Ka
S
HWL.
PSR.
BPC.
Ch.PQ.
K.S.
In.
DS
BRS
B.A.
GEOLOGY
a) COVERING ROCKS:-
SOIL . . . . . S.
KARROO SYSTEM . . . . . Ka.
DOLOMITE SERIES . . . . . D.S.
BLACK REEF SERIES . . . . . B.R.S.
Transvaal System
b) UNITS OF THE KIMBERLEY GROUP:-
MAY REEF HANGINGWALL LEADER . . . . . H.W.L.
SILICEOUS QUARTZITE BAR . . . . . S.Q.B.
PUDDINGSTONE REEF . . . . . P.S.R.
BIG PEBBLE CONGLOMERATE . . . . . B.P.C.
CHERT-PEBBLE QUARTZITE . . . . . Ch.P.Q.
KIMBERLEY SHALES . . . . . K.S.
BIRD AMYGDALOIDAL LAVA . . . . . B.A.
c) INTRUSIVE . . . . . In
BOREHOLES
NIGEL MINE . . . . . K
VOGELSTRUISBULT MINE . . . . . VO.
MARIEVALE MINE . . . . . U
BLOEMENDAL Nº 18 . . . . . B
ENGLISH FEET
100
50
0
100
200
300
400
500

PLATE III

Broadly this description is given irrespective of the regional or local interruptions in the sedimentation which results in the elimination of certain horizons and suites of sediments in some areas. Brief reference is, however, made to these geological breaks.

The units are described in ascending sequence, starting with the so-called Kimberley Shale Zone, which is considered as the upper phase of the Main-Bird Series. The borehole columns shown on Plates III and IV, and Fig. 3 would serve to illustrate the variation along the vertical column, especially in the eastern sector, and the differences in the geology of the two sectors. A system of symbols has been devised indicating the various rock types encountered in the Upper Witwatersrand System and certain "key" horizons in the Kimberley group of sediments. These are shown in Fig. 2.

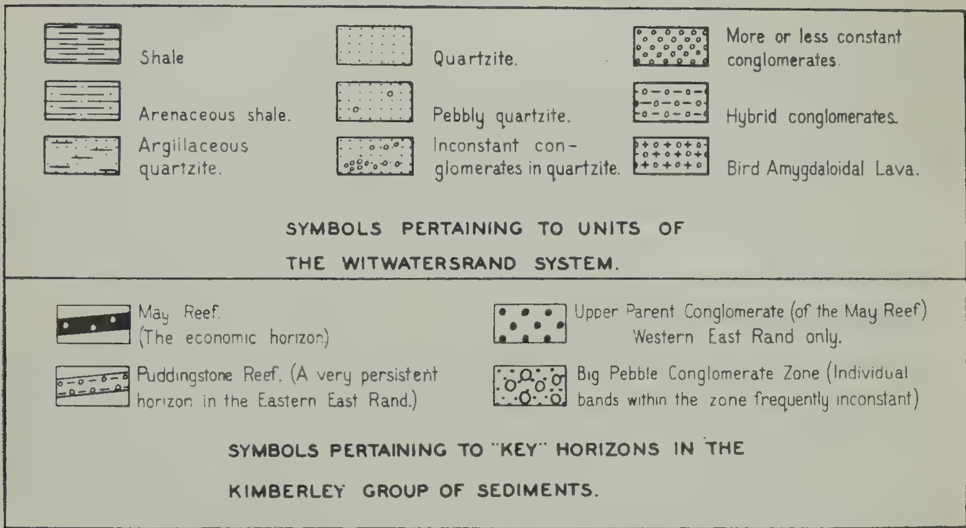


Figure 2

1. *The Kimberley Shale Zone.*

This dominantly argillaceous zone is lithologically very constant. Most thin sections for the microscope were made from the upper layers. Chloritoid has never been observed, but rutile and/or leucoxene are always developed. The rutile is undoubtedly of authigenic origin, occurring as isolated crystals scattered through the ground-mass, or in asteriated clusters, and as bundles or sheaves, exactly as they were shed by the parent titaniferous mineral. A high proportion of the crystals are in the form of geniculate twins.

These shales are easily distinguished from the so-called chloritoid shale a little higher up in the column because of their dark colour (greenish in thin section), parallel banding, absence of any interbedded pebble horizons, absence of chloritoid, and always conclusively if the overlying Chert-pebble Quartzite Zone is present.

On Brakpan Mines and Government Gold Mining Areas the upper phase of the zone is a vitreous, fine-grained quartzite up to 80 feet thick. Generally this upper zone is absent from other parts of the East Rand Basin, apparently owing to erosion before the deposition of the Bottom Reef. Only a few boreholes in the Dunnottar-

Heidelberg area intersected it, while it appears to be totally absent from the eastern sector.

The entire Kimberley Shale Zone is absent from the geological column in some areas, owing to an interruption in the normal sedimentation process: around the May Shaft on the boundary between Daggafontein and East Daggafontein Mines, and in the central portion of the Marievale Mine. The absence of the Kimberley Shales from these areas is attributed to deep erosion before the deposition of the Big Pebble Conglomerate Zone (see Table I).

2. *The Bottom Reef*, which is the basal member of the Kimberley conglomerate group, rests on the Kimberley Shale Zone. It is an inconstant chert-pebble conglomerate, generally only a few inches thick, and is economically unimportant in the East Rand area.

3. *Chert-pebble Quartzite Zone*

This is a zone of pebbly, argillaceous quartzite, with a characteristic greenish colour owing to the presence in the matrix of chlorite and rutile.

Pebble types are well rounded quartz and angular chert, the latter predominating. They are seldom more than one inch in diameter. At various horizons in the zone the pebbles may be sufficiently concentrated to form narrow lenticular conglomerates, sometimes carrying small quantities of gold.

The zone has a maximum thickness of nearly 400 feet, but is absent over appreciable areas, when the Big Pebble Conglomerate comes to rest directly on Kimberley Shales.

4. *The Big Pebble Conglomerate Zone*

Within the East Rand Basin this generally well-developed conglomerate zone varies in thickness between 160 feet and less than two inches, in the latter case sometimes in the form of a fine-grained vitreous quartzite.

Individual pebble bands in the zone tend to change laterally into quartzite, while locally the zone may include lenticular bodies of chloritoid-rich, rutile-rich shale.

Pebble types comprise quartz and chert. Either of the two types may predominate. As the name implies, the pebbles are sometimes very large; occasional chert members measuring up to 9 inches in longest diameter have been encountered. Most of the pebbles are, however, less than 2 inches in diameter.

The sum total of the gold in the zone is sometimes high, but gold is never sufficiently concentrated in any one particular band to make it an economic horizon. The highest values (and the biggest pebbles) tend, however, to occur at the base of the zone.

5. *The Puddingstone Reef*

The mode of origin of this peculiar conglomerate is not clearly understood. Pebbles of various kinds occur scattered in an argillaceous matrix, the latter sometimes approaching a true shale in composition and texture. The matrix has generally, in fact, almost invariably, a characteristic khaki-green colour, which it owes to green chlorite and yellow rutile. It is considered that this conglomerate was originally dark grey in colour owing to the deposition with the pebbles of numerous small grains of black ilmenite. As a result of metamorphism the ilmenite yielded rutile, which is yellow and sometimes foxy-red in colour.

Pebble types comprise chert, quartz, occasional quartzite and isolated red jaspers. An occasional altered igneous rock has been observed. The pebbles are

generally not more than 1½ inches in diameter, although occasional quartzite boulders up to 18 inches in diameter have been encountered. This quartzite compares well with the upper quartzite phase of the Kimberley Shale Zone on Brakpan Mines and Government Gold Mining Areas. It is therefore considered possible that the boulders have been derived from this horizon, from an area eroded contemporaneously.

The Puddingstone Reef is a characteristic rock in the Eastern East Rand, where the overlying rocks are dominantly argillaceous. In this area it apparently represents a transition stage between the Big Pebble Conglomerate below and the "Chloritoid Shale" Marker above. In the Western East Rand where the latter suite of sediments consists chiefly of quartzites and conglomerates, the Puddingstone Reef is seldom recognised in its typical form.

6. *The "Chloritoid Shale" Marker*

What is interesting about this suite of sediments, which may attain a thickness of several hundred feet, is the very marked change of facies from predominantly coarse quartzites and conglomerates to predominantly argillaceous quartzites and fine-grained shales when it is traced from west to east across the East Rand Basin.

The conglomerates in the suite are not always readily distinguishable from other conglomerates in the Kimberley group. Hybrid conglomerates, resembling the Puddingstone Reef in texture and composition, have been observed at intervals within the zone, particularly in the eastern sector. All these conglomerates are as a rule very lenticular in habit.

In the eastern sector flat discoidal shale "pebbles" have been observed amongst the quartz and chert members in the narrow lenticular pebble beds associated with the suite. Locally angular fragments of shale up to two inches long occur as "inclusions" in quartzite and quartzitic shale near the base of the suite. It is obvious that these shale fragments did not travel far, having been derived from a nearby source, apparently from the lower shale phases of their present host.

There is evidence of much scouring and re-deposition during this period. In some areas these deposits may therefore be described as "channel-like", but considering their persistence as a suite over large areas, it is thought that they are not erosion channel fillings in a true sense.

Much ilmenite was also deposited during this period, as evidenced by the presence of a great abundance of undoubted secondary rutile in the form of bundles and sheaves of minute needles and geniculate twins, particularly in the more argillaceous phases. The intense khaki-green shale bands are considered to have been black or dark grey ilmenite muds at the time of deposition.

Chloritoid, well-known as a stress mineral, is quite common in the more argillaceous phases of this zone of sediments, not only in the narrow shale bands, but also in the thick zones of a few hundred feet.

Locally the "Chloritoid Shale" Marker rests directly on the Chert-pebble Quartzite Zone, both the Puddingstone Reef and the Big Pebble Conglomerate then being absent from the section.

7. *"Upper Parent Conglomerate of the May Reef"*.

On Government Gold Mining Areas this zone of conglomerates is known as the U.K.9B Reef. Until now it has been known as the Lower May Reef on the East Rand properties of New Consolidated Gold Fields. It was necessary to drill many boreholes before this conglomerate zone could be placed with certainty (Plate VIII (*b*)). It is now known to be a footwall conglomerate of the May Reef, in actual fact the

principal parent rock of the May Reef in the Western East Rand. It apparently does not exist in the eastern sector.

The Upper Parent Conglomerate is a zone of conglomerates varying in thickness from 2 feet to about 55 feet. Pebbles are almost exclusively of white and dark vein quartz up to 2½ inches in diameter. Individual bands are sometimes well mineralized, but values are generally extremely low. Gold is more or less evenly distributed at an average of about 0·1 to 0·2 dwt. per ton (Plate VII (*b*)). On Government Gold Mining Areas the zone is slightly richer in gold. Isolated high values have been encountered, but these are very sporadic, and not confined to any particular band. Very high silver values have been recorded on Government Gold Mining Areas, where it has been found necessary to have all samples parted for silver in assaying.

Locally, an ill-sorted reef carrying sporadic payable gold values occurs along the base of the zone. This Sharpe (1942-1945 and 1949) calls the U.K.9C Reef.

8. *Micaceous-gritty Quartzite*

The Upper Parent Conglomerate is overlain by a peculiar, sometimes highly micaceous quartzite, which also, is only encountered in the Western East Rand. It is up to 20 feet thick, and forms the immediate footwall of the May Reef over large areas.

9. *The May Reef*

In the Western East Rand the May Reef comes to rest alternately on the Micaceous-gritty Quartzite and the Upper Parent Conglomerate. The relationship is one of a basal conglomerate blanketing an erosion surface that truncates wide open anticlines and synclines of small amplitude. The angular difference is therefore small, amounting at the most to a few degrees. The coarse and heavy fraction of the eroded portions of the Upper Parent Conglomerate was concentrated, and finally deposited largely on the sub-May Reef synclines (Plate VIII and Fig. 6).

In the western sector the May Reef is generally very thin, seldom exceeding 24 inches, although greater widths up to 5½ feet are known locally. The highest values are generally limited to the upper portion of the reef; most of the gold is frequently concentrated in the uppermost few inches. It is a small-to-medium-pebble conglomerate, containing exclusively vein quartz pebbles which seldom exceed one inch in diameter. In this sector the May Reef is of economic importance on Government Gold Mining Areas.

The May Reef is strikingly unconformable to its footwall beds in the Eastern East Rand. The older gravels (i.e. Puddingstone Reef, Big Pebble Conglomerate, etc.) furnished the material, including the gold, for the formation of the May Reef, as a result of pre-May Reef folding and uplift, with consequent erosion to a base level of deposition. The more robust bodies of May Reef are generally found on the sub-May Reef synclines.

In this sector the May Reef varies rapidly from place to place, ranging in width from a mere contact to a robust body of up to 10 feet. Pebbles are generally of quartz and chert. The latter may be very large, up to 7 inches in longest measurement.

A feature of great interest and economic importance in the eastern sector is the occurrence near the top of the May Reef of a thin layer of fine-grained material rich in gold, metallic sulphides and other heavy minerals, such as chromite and zircon (detrital), and rutile which is clearly secondary in origin. This thin layer generally indicates high gold values, and in borehole cores it is of great diagnostic value. Sampling frequently shows the highest concentration of gold towards the top of the reef (Plate V 4 (*c*)).

10. *Siliceous Quartzite Bar*

In the Western East Rand the May Reef is invariably overlain by a pale, vitreous fine-grained quartzite up to 13 feet in thickness. It is a very important marker horizon in this area.

In the eastern sector the bar is usually much thinner, and is absent over appreciable areas. It is frequently bluish-black in colour.

11. *The May Reef Hangingwall Leader*

This conglomerate horizon exists from one inch to 13 feet above the May Reef in the eastern sector. It has not been encountred in the western sector.

It varies from a few scattered pebbles in a vitreous quartzite to a robust conglomerate up to 9 feet wide, with vein quartz pebbles of about one inch in diameter predominating. It is not a component part of the May Reef, although it may sometimes be extremely difficult to separate the two horizons.

Apart from a fluctuation in thickness, it maintains its characteristic appearance throughout the eastern area, whereas the May Reef immediately below exhibits a marked lateral change in pebble types and sizes, and the mineralization of the matrix.

Around the No. 2 Shaft on Vogelstruisbult the May Reef Hangingwall Leader has come to rest directly on the "Chloritoid Shale" Marker, i.e. the May Reef does not exist in this area (Plates V 3 (*b*) and VII).

The May Reef Hangingwall Leader generally has a low gold content, which assists in determining the upper contact of the May Reef, when the two conglomerate bands occur close together. On East Daggafontein Mine, however, a narrow band carried payable quantities of gold over short distances.

12. *Argillaceous Quartzite Bar*

On the Vogelstruisbult Mine a thin film of argillaceous material usually occurs directly above the May Reef Hangingwall Leader. In stoped areas the rocks are seen to part along this plane. Frequently it causes trouble during stoping operations, making it difficult to control the stoping width. Locally this thin layer of argillaceous material swells out into an arenaceous shale or argillaceous quartzite up to 15 feet thick. It now contains much chloritoid and rutile, and may be indistinguishable from some of the May Reef footwall rocks. In one instance on Vogelstruisbult, beyond a fault, a haulage was erroneously directed upwards for a short distance in this bed. It is not known in the western sector.

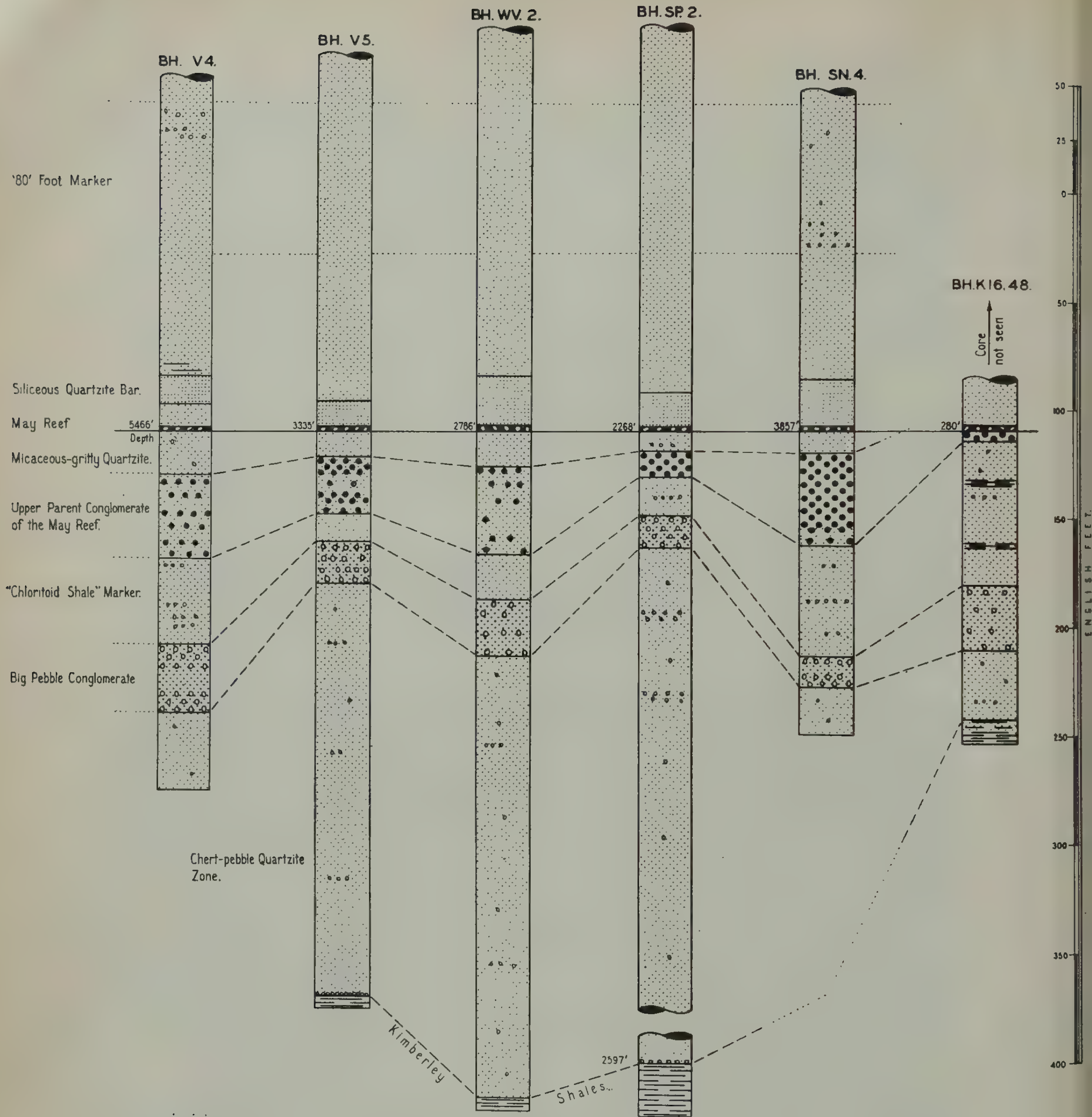
13. *Coarse-grained Quartzite* (*Sharpe's U.K.*8)

Following above the Argillaceous Quartzite Bar in the Eastern East Rand, and above the Siliceous Quartzite Bar in the Western East Rand, is a zone of coarse-grained quartzite varying in thickness between 30 and 100 feet. Measurement of the true thickness of the zone is seldom possible, because the lower limit of the "80" Foot Marker is usually indefinite.

14. *The* "80" *Foot Marker*

This is a conglomerate zone which varies in thickness from nearly 100 feet on Brakpan Mines to less than 5 feet in portions of the Eastern East Rand. Southwards towards Heidelberg the zone deteriorates so much, that boreholes frequently fail to disclose its existence.

As most of the individual conglomerate bands are short-lived, it is often impossible to measure the thickness of the zone (Plates III and IV). The uppermost



WESTERN EAST RAND.

COMPARATIVE KIMBERLEY BOREHOLE SECTIONS.

(ARRANGEMENT ARBITRARY.)

VLAKFONTEIN MINE..........V
WEST VLAKFONTEIN MINE....WV.
SPAARWATER MINE....SP.
SUB NIGEL MINE.....SN.
NIGEL MINE:.........K.

PLATE IV

band is the most persistent, and on Sub-Nigel, Vogelstruisbult, Nigel and Marievale Mines, it frequently lies at $\pm$ 80 feet vertically above the May Reef. On Government Gold Mining Areas the upper band exists up to 150 feet above the May Reef, and locally contains payable quantities of gold.

Locally, near the sub-outcrop against the Karroo System, the zone contains numerous grains and small pebbles of pyrophyllite, scattered amongst the quartz and chert members. The mineral is very soft and can be scratched by a finger nail. Small pebbles of chlorite, although not as abundant as pyrophyllite, are more widely distributed along this horizon.

15. *Groups of quartzites and groups of conglomerates*, also belonging to the Kimberley beds, exist higher up in the column. These conglomerates are economically unimportant.

Towards the northern extremity of the western sector the highest conglomerate band belonging to the Kimberley group occurs at about 800 feet above the May Reef. Eastwards and south-eastwards the section thins appreciably, until in Vogelstruisbult No. 4 Shaft the highest band lies at a distance of 480 feet above the May Reef.

A section from Brakpan Mines to East Daggafontein (Whiteside, 1950) shows that this marked thinning is not the result of any interruptions in the sedimentation, but takes place along each individual quartzite and conglomerate zone. In the Dunnottar-Heidelberg area these upper conglomerate clusters are poorly developed, having changed into quartzites near Heidelberg.

16. *The Kimberley-Elsburg Intermediate Quartzites*

These are more or less clear quartzites separating the Kimberley group of conglomerates from those of the Elsburg group (Plates I and II).

In Sub-Nigel No. 3 Shaft the zone is apparently 1,388 feet thick. In this shaft the three major sedimentary groups under consideration have the following apparent thicknesses:

Elsburg conglomerate group	700	feet
Kimberley-Elsburg Intermediate Quartzites ...	1,388	,,
Kimberley conglomerate group	876	,,
Total from base of Ventersdorp Lava to Kimberley Shales	2,964	feet

In Vlakfontein No. 1 Shaft the corresponding thicknesses are 625, 1,682, 922 and 3,229 feet.

17. *The Elsburg Conglomerate Group*

The average thickness of the group in the East Rand Basin is 700 feet. Individual conglomerates are often very robust, but generally impersistent. Considered as a whole they appear to be more lenticular than most members of the Kimberley group. No economic horizons have been discovered amongst the Elsburg reefs in the East Rand Basin, their gold content being without exception very low.

VI THE STRATIGRAPHY OF THE KIMBERLEY GROUP OF SEDIMENTS IN THE EASTERN HALF OF THE BASIN.

This sector includes the eastern half of Springs Mines, Daggafontein Mine, East Daggafontein Mine, the bigger portion of the Vogelstruisbult Mine, Marievale Mine, farm Bloemendal 19 and the eastern half of the Nigel Mine (Plate VII).

The exceptional variation in the stratigraphic column could be studied in great detail in this area, because of the large number of boreholes drilled from surface, and the extensive underground development carried out on Vogelstruisbult, Daggafontein and East Daggafontein.

The stratigraphy of the Kimberley group of sediments with particular reference to the Kimberley Economic Zone will be described here for the different areas as enumerated above. This cannot be done without reference to the three regional erosion surfaces, namely: immediately below the Bottom Reef, Big Pebble Conglomerate and the May Reef. Evidence will also be produced of local scour at other points along the stratigraphic column in many localities within this area. The measured thicknesses of the major suites of sediments below these erosion surfaces are thus seldom the original depositional thicknesses, although these could also vary within wide limits. An erosion surface is therefore not proved, and not necessarily indicated, by the varying thickness of a particular sedimentary suite. Detailed stratigraphic study is necessary in order to determine whether or not any formations have been eliminated.

Differential uplift of the depositional floor and erosion to a base level earlier than the deposition of the Bottom Reef, apparently took place over a large area, resulting in the elimination of the upper quartzite phase of the Kimberley Shale Zone in such areas of uplift.

Evidence will be produced of a period of erosion that preceded the deposition of the Big Pebble Conglomerate Zone. This erosion surface generally has a low relief, but in some areas there is evidence of scour to some considerable depth, resulting in the elimination of the entire Kimberley Shale Zone (Fig. 3). Renewed deposition appears to have been continuous across these areas of deep scour, resulting in a considerable thickening of the individual units as compared to neighbouring areas (Fig. 3).

The floor of the May Reef could be mapped in the mine workings (Plate VI). This erosion surface always has a low relief.

These three regional unconformities will be described in more detail in later chapters, and their economic and geological significance will be discussed.

As the lithology of the various horizons has been described in some detail in Chapter V, all the details will not be repeated here. Additional information will be given, and the inter-relationships of the various units will be discussed.

Some of the geological sections that accompany this description (Plate V, 1, 2 and 3) have been plotted below a common datum, which is an imaginary plane 6,000 feet above sea level.

A. Eastern Two-thirds of the Vogelstruisbult Mine

The area around No. 3 Shaft falls in the western sector.

No. 1 and No. 2 Shafts penetrated the Kimberley group of sediments and proceeded down to the Main Reef. No. 4 Shaft was sunk down to the May Reef only, while 12 boreholes were drilled from surface, of which one (Vo. 1) was stopped in the Kimberley Shales without the May Reef being identified. The other boreholes, Vo.3 to Vo.13, drilled in the outcrop area, all intersected the May Reef (Plate VII).

Extensive development on the May Reef has been carried out east of No. 1 Shaft to beyond No. 4 Shaft. No. 2 Shaft is connected with this development by a single haulage.

Numerous boreholes have been drilled from the underground workings, not only to determine the May Reef position and value beyond faults, but also the inter-relationships of the various horizons, and the nature of the sub-May Reef erosion surface.

(*a*) In the No. 1 Shaft the Kimberley Shale Zone is 356 feet thick, including 65 feet quartzitic shale near the top of the zone.

In No. 2 Shaft the zone is only 247 feet thick, the upper 114 feet being a quartzitic shale.

These quartzitic phases are, however, not the equivalent of the upper vitreous quartzite phase as known in the Brakpan area.

(*b*) The Chert-pebble Quartzite Zone is a very useful marker in locating the May Reef beyond the faults in the mine. The thickness of the zone varies from 40 to 120 feet, the thicker developments occurring east of No. 1 Shaft. Beyond faults the zone can be confused with a lithologically similar pebbly zone associated with the "80" Foot Marker (Plate V 3 (*a*) and Fig. 5).

(*c*) The Big Pebble Conglomerate is of blanket character and is a very useful marker. It varies in thickness from a few inches to a maximum of about 18 feet. It is generally a well-packed conglomerate, but locally portions of the body change laterally into quartzite (Plate V 3, and Plate XI, Ph. 10). Locally it may also include lenses of chloritoid-rich, rutile-rich shale.

The sum total of the gold in the zone is generally high in the area east of No. 1 Shaft. Spectacularly high values have been intersected on the bottom bands, but these values never occur over long distances. The zone becomes poorer in gold southwards towards No. 2 Shaft and the outcrop area.

The chert cobbles are usually of diagnostic value, although locally the May Reef includes similar cobbles.

In some localities the May Reef horizon transgresses on to the Big Pebble Conglomerate. The erosion surface may also cut right through this horizon. In these areas the May Reef usually deteriorates into a pencil-line contact.

Locally the Big Pebble Conglomerate is also absent, apparently owing to scour before the deposition of the Puddingstone Reef or "Chloritoid Shale" Marker.

(*d*) The Puddingstone Reef is a very persistent horizon in this area, and because of its characteristic and constant lithological character, it is the most useful marker to all those engaged in mining the May Reef. This peculiar hybrid conglomerate with its argillaceous matrix varies in thickness from a few inches to a maximum of about 15 feet. Locally it includes lenses of chloritoid-rich, rutile-rich shale. Its gold content is generally very low, although locally it may contain fair values (Plate V, 4 (*b*)).

Physically it appears to be a transitional phase between the Big Pebble Conglomerate below and the rutile-rich shales above. Locally it is absent from the column, apparently as the result of erosion before the deposition of the "Chloritoid Shale" Marker, and over bigger areas as the result of the period of erosion that gave rise to the formation of the May Reef (Plate VI).

(*e*) The "Chloritoid Shale" Marker is perhaps the most interesting suite of sediments in the Kimberley-Elsburg Series, because the beds exhibit a very marked change of facies regionally, often perceptible over very short distances. There is also evidence of a certain amount of scouring and re-deposition during this period.

In the area under review these beds are dominantly argillaceous, locally including quartzite zones indistinguishable from the May Reef hangingwall quartzites. Inconstant conglomerates occur at intervals through the zone (Plate V, 3 (*b*).

On Vogelstruisbult the suite has a maximum thickness of 177 feet in No. 2 Shaft, gradually thinning eastwards. From east of No. 1 Shaft to beyond No. 4 Shaft it has a maximum thickness of 50 feet, but is absent over appreciable areas, as a result of the sub-May Reef period of erosion (Plates V, 1, 2, 3 and VI).

In No. 2 Shaft the suite can be divided into an upper shale division, a middle quartzite division and a lower shale division. Eastwards along 2K Haulage (Plate V, 3 (*b*)), where a large number of boreholes has been drilled through the zone, this condition still holds in some of the boreholes, but the change is mostly rapid, and possibly abrupt in some instances. The quartzites of the zone intersected by borehole 398 are indistinguishable from the May Reef hangingwall quartzites. Along this haulage an attempt has been made to classify the sediments of the suite into quartzites, argillaceous quartzites, arenaceous shales, shales, pebbly shales (borehole 415), and lenticular conglomerates.

East of No. 1 Shaft the same general lithological description would apply, except that the quartzite units are thinner, and that the upper shale phase is apparently everywhere eliminated.

In the outcrop area on Vogelstruisbult (boreholes Vo.3 to Vo.13) the zone varies in thickness from 20 to 160 feet in a very limited area (Plate VII). In some of these boreholes it is possible to recognise the middle quartzite and the lower shale divisions of No. 2 Shaft, but in most (e.g. Vo.11, Plate III) the change along the vertical column is very impressive. The different units are, however, perfectly transitional to one another, sometimes exhibiting also a transitional relationship with the Puddingstone Reef below (e.g. Vo.9, Plate III).

The angular shale fragments that occur locally as "inclusions" in quartzite and quartzitic shale, have obviously been derived from a source near by, apparently from the lower shale phase of the suite, and incorporated, without being destroyed, in the succeeding layers. That would indicate a break in the sedimentation near the bottom of the zone in some areas (Plate II).

Sharpe (1942-1945 and 1949) regards this group of sediments as typical channel deposits, as the infilling of narrow longitudinal hollows cut into the underlying rocks. But that does not seem to be entirely correct, as regionally the basal members of the group appear to have been deposited on a fairly even floor. The marked lateral variation creates the impression that these beds are true channel fillings. Furthermore, they have very wide distribution, although individual lithological units are frequently highly lenticular. They have been proved to be absent only in some well-defined areas in the eastern sector as the result of pre-May Reef folding and uplift, with erosion to a base level of deposition as a sequel, and very locally in the western sector as a result of erosion before the deposition of the Upper Parent Conglomerate of the May Reef. Elsewhere their distribution is controlled by the outcrop and sub-outcrop of the bottom surface of the zone.

(*f*) The May Reef, which apparently is the only economic horizon amongst the Kimberley reefs in the East Rand Basin, has certain characteristic features. Primarily, as mentioned before, it lies unconformably on the older Kimberley beds.

In the area under review the May Reef horizon varies in thickness from a pencil-line contact to a robust body 10 feet wide. It exhibits a marked lateral change. In places it is a narrow small-pebble conglomerate containing smoky quartz pebbles only. In adjacent areas it develops into a robust large-pebbled conglomerate con-

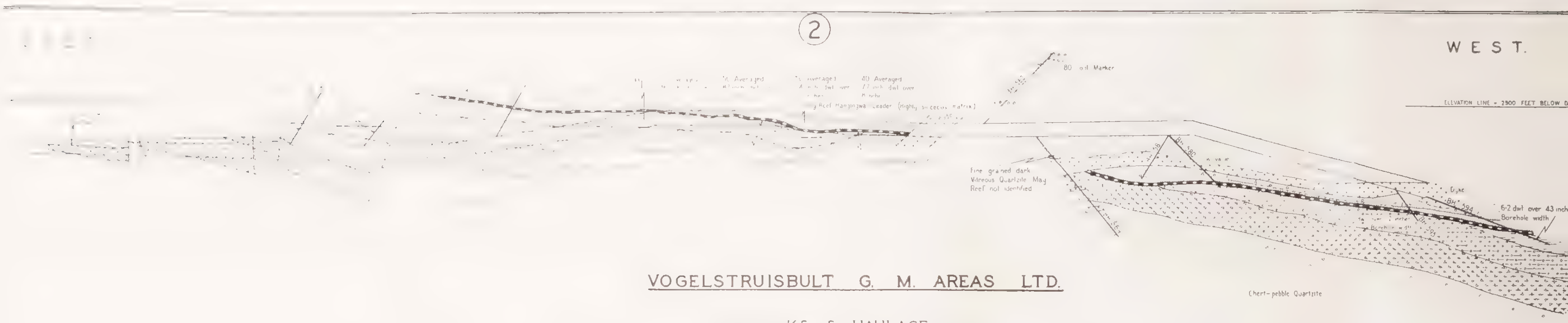
(1)

VOGELSTRUISBULT MINE
EASTERN EAST RAND

MEASURED SECTIONS ALONG DEVELOPMENTS

MAY REEF HANGINGWALL LEADER	H.W.L.
CHLORITOID SHALE	Ch.Sh.
PUDDINGSTONE REEF	P.S.R.
BIG PEBBLE CONGLOMERATE	B.P.C.
FAULT	

(2)



VOGELSTRUISBULT G. M. AREAS LTD.

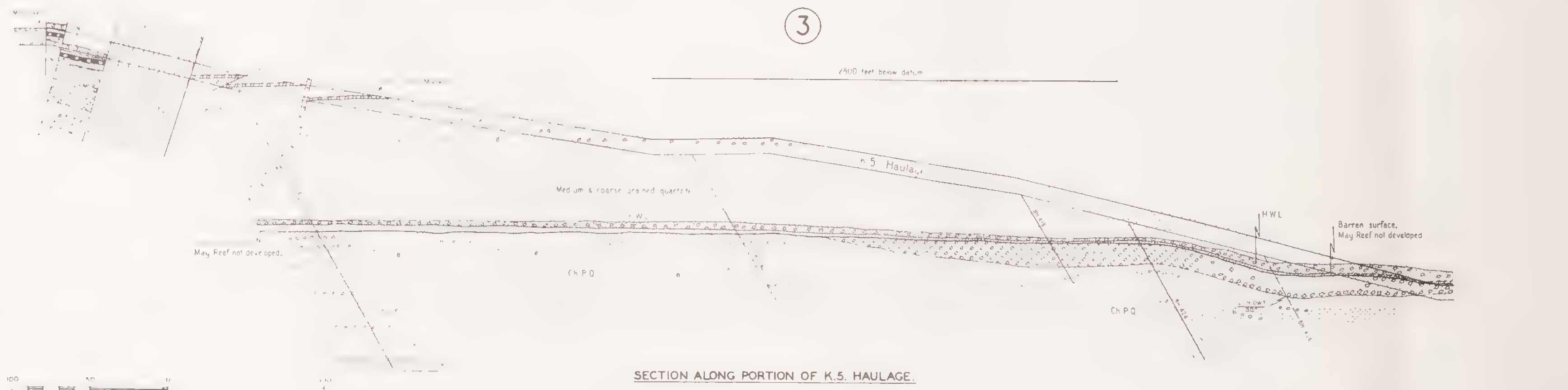
K5 5 HAULAGE

GEOLOGICAL SECTION ILLUSTRATING ANGULAR UNCONFORMITY BELOW THE MAY REEF.

Modified after F.S.J. de Jager Trans. & Proc. geol. Soc. S.Afr. Vol. LII 1949.

(3)

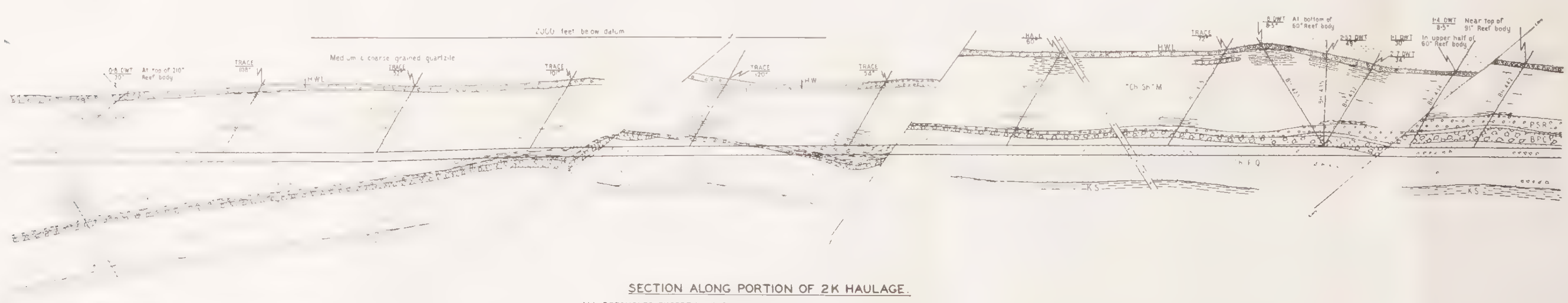
a)



SECTION ALONG PORTION OF K.5. HAULAGE.

ALL BOREHOLES, EXCEPT No 438, LIE APPROXIMATELY IN THE SAME VERTICAL PLANE

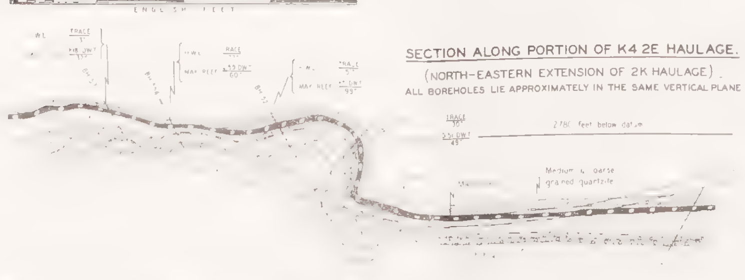
b)



SECTION ALONG PORTION OF 2K HAULAGE.

ALL BOREHOLES, EXCEPT No 435, LIE APPROXIMATELY IN THE SAME VERTICAL PLANE

c)



SECTION ALONG PORTION OF K4 2E HAULAGE.

(NORTH-EASTERN EXTENSION OF 2K HAULAGE)

ALL BOREHOLES LIE APPROXIMATELY IN THE SAME VERTICAL PLANE

MAY REEF HANGINGWALL LEADER	H.W.L.
CHLORITOID SHALE MARKER	Ch.Sh.M.
CHLORITOID SHALE	Ch.Sh.
PUDDINGSTONE REEF	P.S.R.
BIG PEBBLE CONGLOMERATE	B.P.C.
CHERT-PEBBLE QUARTZITE	Ch.P.Q.
KIMBERLEY SHALES	K.S.
FAULTS	
DYKES	
GOLD IN DWT PER TON / APPARENT WIDTH IN INCHES	= 2·33 DWT / 49"

VOGELSTRUISBULT MINE – EASTERN EAST RAND

SECTIONS DETERMINED BY DRILLING

NOTE – IN ORDER TO SHOW DETAIL, VERTICAL SCALE SLIGHTLY EXAGGERATED IN PLACES

a) K4 1B DRIVE

(4)

b) K4 1B3 RAISE

MAY REEF HANGINGWALL LEADER	H.W.L.
CHLORITOID SHALE	Ch.Sh.
PUDDINGSTONE REEF	P.S.R.
BIG PEBBLE CONGLOMERATE	B.P.C.
MAY REEF VALUE: 4 / 172	SAMPLE WIDTH IN INCHES / DWT/TON.
FAULTS	

c) K4 1 HAULAGE

VOGELSTRUISBULT MINE – EASTERN EAST RAND

MEASURED SECTIONS ALONG DEVELOPMENTS

ENGLISH FEET

taining large quartz pebbles and cobbles of chert, the latter up to 7 inches long. It may consist of several bands, separated by bands of fine-grained vitreous quartzite.

Pebble types in the May Reef comprise black and banded cherts, white-vein quartz (frequently with dark rims), dark smoky quartz and occasional opalescent blue quartz. In one instance only was the May Reef observed to include a 1¼-inch pebble of khaki-green argillaceous quartzite, lying about two inches above the bottom contact of the reef. At this point, and in the immediate vicinity, the May Reef rests on a similar rock.

The matrix of the reef is as a rule very dark in colour. Characteristic of the highly payable May Reef is its "ashy" matrix, consisting of carbon, pyrite, pyrrhotite, rutile, gold, zircons, chromite, etc. In one instance a ¼-inch layer of brittle carbon traversed by numerous thin veinlets of gold was observed along the bottom contact of a 4-inch reef. The pyrrhotite usually takes the form of thin flakes, frequently in fractures in the pebbles. A member of the Sampling Department found a large crystal of sphalerite on the bottom contact of the reef.

As mentioned in Chapter V, a thin layer of fine-grained heavy minerals frequently occurs near the top of the May Reef in rich areas. It includes the bulk of the gold and other heavy minerals found in the May Reef. Under the microscope it was possible to distinguish between the allogenic and authigenic constituents of this thin undoubtedly sedimentary layer:

1. *Primary or Allogenic Constituents:*

(i) *Chromite*

The concentrate is very rich in chromite. Most of the grains are well-rounded and up to 0.3 mm. in diameter.

In thin section the chromite grains become translucent and appear brown by transmitted light. Some of the grains are fractured, apparently *in situ*, the cracks having afterwards been filled with pyrite.

(ii) *Zircon*

This mineral appears in the form of well-rounded grains, up to 0.3 mm. in diameter. Although always present, it is not as abundant as chromite.

(iii) Detrital quartz grains.

(iv) A very high concentration of gold, the bulk of which is apparently now locked up in the pyrite and pyrrhotite.

2. *Secondary or Authigenic Constituents:*

(i) "Buckshot" pyrite up to 0.5 mm. in diameter. Numerous crystal faces can be observed in the small bodies. Occasionally the pyrite is surrounded by dark rims, possibly caused by carbon. Around their edges the small pyrite bodies are seen to replace detrital quartz grains.

Stringers of pyrite occur in the matrix and may cut across the quartz grains.

(ii) Very small crystals of rutile are always present, usually in bundles and sheaves. Many of the crystals are twinned. It is believed that the rutile betrays the former presence of detrital ilmenite.

(iii) Chloritoid containing abundant dark inclusions.

(iv) Sericite and chlorite.

The pyrrhotite is generally macroscopic.

Around No. 2 Shaft and eastwards along 2K Haulage for a distance of nearly 6,000 feet, the May Reef Hangingwall Leader comes to rest directly on the "Chloritoid Shale" Marker, possibly eliminating a pre-existing May Reef (Plate V, 3 (*b*) and (*c*), and Plate VII). This condition could be explained by slight post-May-Reef warping of the depositional floor, resulting in the May Reef being stripped off from the higher areas and deposited nearby. Alternatively this area might have stood slightly above the general level of deposition of the May Reef.

(*g*) The "80" Foot Marker consists of a zone of pebbly quartzite 30 feet thick, capped by a well-developed persistent conglomerate band up to 3 feet wide, the latter averaging about 80 feet vertically above the May Reef. The minimum distance is about 60 feet and the maximum 90 feet.

A feature of this conglomerate cluster is the high proportion of chert pebbles; some individuals appear to be of chlorite.

The "80" Foot Marker is invaluable in unravelling structural problems. The sub-May-Reef geology in the area north of the east-west tear fault (Plate VI) was largely determined by making use of this cluster as a datum, especially during the initial stages of development in that area. It took a long time before the May Reef horizon was identified in the boreholes drilled from K4. 1B. 2 Winze (Fig. 5).

B. Marievale Mine

Less than two-thirds of this mine is underlain by the May Reef. The sub-outcrop against the base of the Transvaal or Karroo Systems trends due east-west, with the reef dipping at low angles to the north over the larger portion of the property. But in the western part of the mine the Witwatersrand beds are folded into an asymmetrical anticline, resulting in the sub-outcrop making a sharp horse-shoe bend to the south-east, the reef now dipping at steep angles to the south-west (Plate VII).

Altogether 13 boreholes from surface intersected the May Reef. Two boreholes, UR and UP, drilled not far behind the sub-outcrop of the reef, intersected the lower portion of the Kimberley Economic Zone.

A cross-cut haulage from No. 1 Shaft, which is situated behind the sub-outcrop, was begun from a point just below the Main Reef. It intersected the May Reef some 4,000 feet further to the north, after which a drive, advancing both eastwards and westwards, has now nearly crossed the mine.

In the western part of the mine borehole U.T. is the most instructive (Plate III and Fig. 3): the Kimberley Shale Zone is overlain by 54 feet of Chert-pebble Quartzite, followed by the Big Pebble Conglomerate. The Puddingstone Reef is absent from the section. The "Chloritoid Shale" Marker is 204 feet thick, being a fine-grained chloritoid-rich shale in the upper half, and a sub-glassy quartzite in the lower half. The two sub-divisions are transitional to each other. The May Reef was intersected very near the sub-outcrop against the base of the Transvaal System. The highest concentration of gold occurs in association with the thin layer of fine-grained heavy minerals near the top of the reef.

The May Reef has the chloritoid-rich shales as footwall over large areas in this mine, but it is possible that it rests directly on Kimberley Shales near the common boundary with the farm Bloemendal (boreholes U.X, U.V. and U.J., Plate VII). These boreholes were not drilled deeply enough for correlation purposes.

On Bloemendal the May Reef rests on Kimberley Shales in boreholes B.1., B.2 and B.3 (Plate III and Fig. 3). The quartzites above the shales are typical of the May Reef Hangingwall, and it is impossible to confuse them with the Chert-pebble

K4 1 HAULAGE
EAST DAGGAFONTEIN
K4 2 HAULAGE
KS 3 WINZE
HAULAGE
VOGELSTRUISBULT MINE — EASTERN EAST RAND
PORTION OF KIMBERLEY REEF WORKINGS

PLATE VI

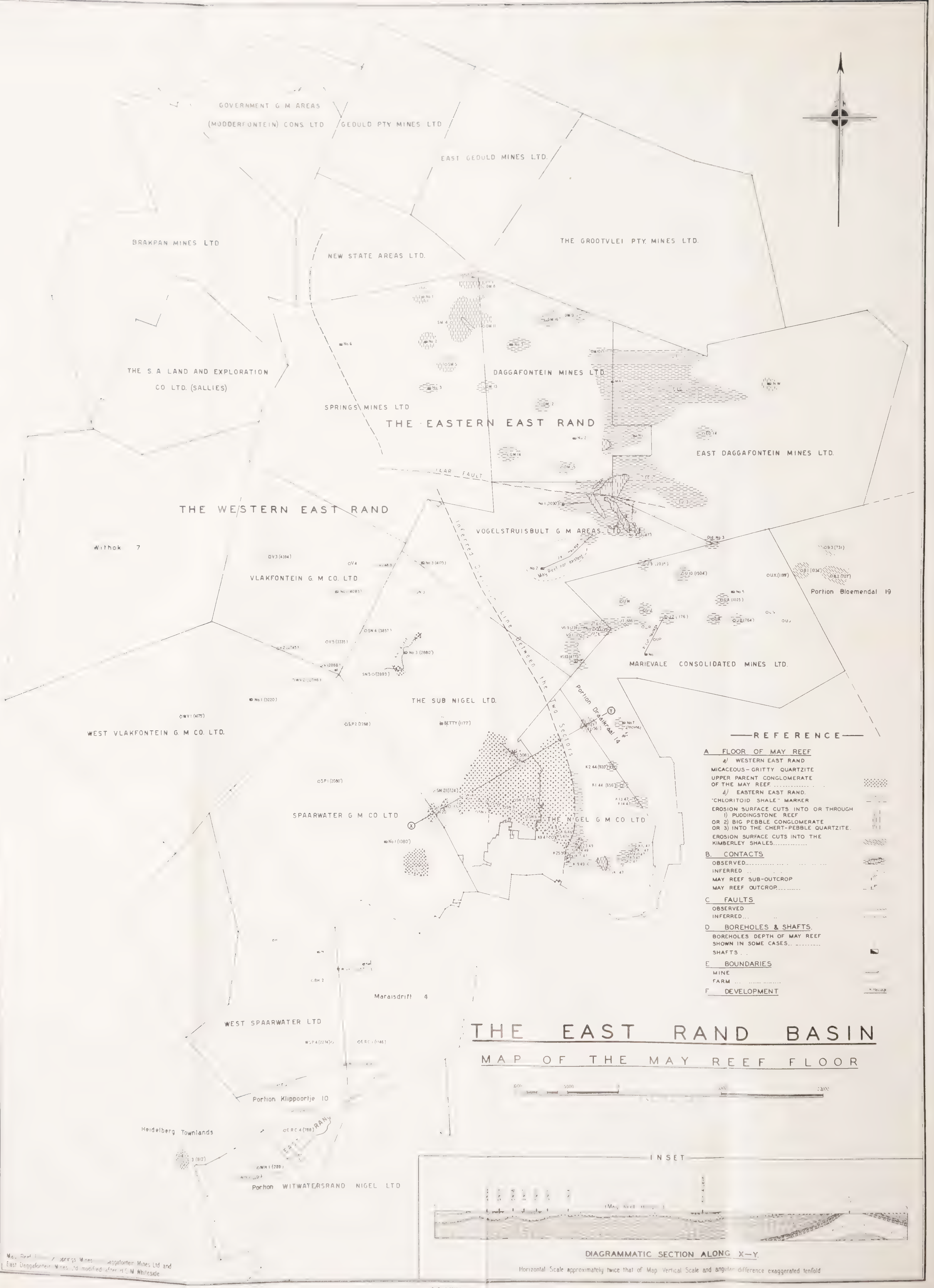

GOVERNMENT G.M. AREAS
(MODDERFONTEIN) CONS. LTD.
GEDULD PTY. MINES LTD.
EAST GEDULD MINES LTD.
BRAKPAN MINES LTD.
NEW STATE AREAS LTD.
THE GROOTVLEI PTY. MINES LTD.
THE S.A. LAND AND EXPLORATION
CO. LTD. (SALLIES)
DAGGAFONTEIN MINES LTD.
SPRINGS MINES LTD.
THE EASTERN EAST RAND
EAST DAGGAFONTEIN MINES LTD.
THE WESTERN EAST RAND
VOGELSTRUISBULT G.M. AREAS LTD.
Withok 7
VLAKFONTEIN G.M. CO. LTD.
Inferred Line Between the Two Sectors
Portion Bloemendal 19
MARIEVALE CONSOLIDATED MINES LTD.
THE SUB NIGEL LTD.
Portion Droogekraal 14
WEST VLAKFONTEIN G.M. CO. LTD.
SPAARWATER G.M. CO. LTD.
THE NIGEL G.M. CO. LTD.
Maraisdrift 4
WEST SPAARWATER LTD.
Portion Klippoortje 10
Heidelberg Townlands
Portion WITWATERSRAND NIGEL LTD.
REFERENCE
A. FLOOR OF MAY REEF
a) WESTERN EAST RAND
MICACEOUS-GRITTY QUARTZITE
UPPER PARENT CONGLOMERATE OF THE MAY REEF
b) EASTERN EAST RAND
"CHLORITOID SHALE" MARKER
EROSION SURFACE CUTS INTO OR THROUGH 1) PUDDINGSTONE REEF OR 2) BIG PEBBLE CONGLOMERATE OR 3) INTO THE CHERT-PEBBLE QUARTZITE
EROSION SURFACE CUTS INTO THE KIMBERLEY SHALES
B. CONTACTS
OBSERVED
INFERRED
MAY REEF SUB-OUTCROP
MAY REEF OUTCROP
C. FAULTS
OBSERVED
INFERRED
D. BOREHOLES & SHAFTS
BOREHOLES DEPTH OF MAY REEF SHOWN IN SOME CASES
SHAFTS
E. BOUNDARIES
MINE
FARM
F. DEVELOPMENT
THE EAST RAND BASIN
MAP OF THE MAY REEF FLOOR
INSET
DIAGRAMMATIC SECTION ALONG X-Y
Horizontal Scale approximately twice that of Map. Vertical Scale and angular difference exaggerated tenfold

PLATE VII

Quartzite Zone. Furthermore, the shales intersected by these three boreholes are typical of the Kimberley Shales, having the following diagnostic features:

(1) dark, fine-grained with parallel banding,
(2) absence of any pebbly layers,
(3) chloritoid did not develop,
(4) transitional to the Main-Bird quartzites below.

In the extreme south-western portion of Marievale near the common boundary with the Nigel Mine, borehole U.Y. showed the May Reef to lie directly on the Chert-pebble Quartzite Zone. Around this borehole the sub-May Reef erosion surface has thus eliminated the principal footwall conglomerates of the May Reef (Plates III and VII).

Three boreholes, U.A., U.P. and U.R., in the central portion of the mine, deserve special attention because the Kimberley Shales are absent from these sections (Fig. 3).

Borehole U.A.

The May Reef lies on dark, fine-grained chloritoid-rich shale. Including a few thin dykes or sills, the zone has a borehole thickness of 336 feet. Chloritoid developed throughout the zone. Two distinctly different conglomerate beds occur in contact with each other at the base of this shale zone. The upper one, 18 feet thick, is similar in all respects to the Puddingstone Reef as known elsewhere in the East Rand Basin, and the lower horizon, 8 feet thick, looks exactly like the Big Pebble Conglomerate in its typical form. The lower horizon is separated from the Bird Amygdaloid by 50 feet of clear quartzites and thick intrusions.

Borehole U.R.

This borehole was drilled behind the May Reef sub-outcrop (Fig. 3). The Black Reef Quartzite is underlain by 456 feet of predominantly argillaceous rocks, which can be divided into an upper shale division rich in chloritoid, a middle division of sub-glassy quartzite containing conglomerate bands in which angular chert pebbles predominate, and a lower shale division in which chloritoid was not observed.

The above zone is underlain by a conglomerate zone 112 feet thick, similar in all respects to the Big Pebble Conglomerate as known elsewhere. The zone is a well-packed conglomerate in the lower half, consisting of quartz and chert pebbles up to 2 inches in diameter, but it grades upwards into a quartzite. It rests on the Bird Amygdaloid.

Borehole U.P.

This borehole was also drilled behind the May Reef sub-outcrop. A conglomerate zone, 143 feet thick, and similar in all respects to the Big Pebble Conglomerate, rests on the Bird Amygdaloid. It includes 35 feet of dark chloritoid-rich shale in its lower half. It is overlain by a zone of argillaceous quartzite, 166 feet thick, which contains a little chloritoid and much secondary rutile. A 50-feet-thick intrusion divides this zone.

The cross-cut haulage from No. 1 Shaft revealed a section similar to those described above, except for 30-feet typical Kimberley Shale intervening between the robust conglomerate zone and the Bird Amygdaloid.

It was originally considered that a deep erosion channel underlay the May Reef here, filled with gravels in the lower half, and with muds and sands in the upper half. But evidence against an infilled erosion channel thus dated, is the fact that the litho-

logical sequence immediately below the May Reef appears to be similar in all respects to the stratigraphic sequence immediately below the May Reef in adjacent areas, where the geological column is more or less complete. The author therefore prefers to correlate the robust conglomerate zone described above with the Big Pebble Conglomerate, and the succeeding predominantly argillaceous zone with the "Chloritoid Shale" Marker—compare borehole U.T. further west, and borehole Vo.11 in the outcrop area on Vogelstruisbult (Plate III and Fig. 3).

It is therefore thought that deep scouring took place in this area during the regional erosion interval that preceded the deposition of the Big Pebble Conglomerate, resulting in the elimination of the entire Kimberley Shale Zone and the upper portions of the Bird Amygdaloid. Renewed deposition appears to have been continuous across this low-lying area, resulting in a marked thickening of the Big Pebble Conglomerate and "Chloritoid Shale" Marker as compared to neighbouring areas (Fig. 3).

C. Eastern Half of the Nigel Mine

A large number of boreholes has been drilled along the eastern boundary of the mine, and underground development on the May Reef is in progress near the southern extremity of the area under review, so that a fairly comprehensive picture of the geology can be formed (Plate VII).

(*a*) *The Strip along the Eastern Boundary:*

Lithologically the average section here compares well with the average column on Vogelstruisbult. Excluding borehole U.Y. on the Marievale side of the boundary (Plate VII), there is no further evidence that the sub-May-Reef erosion surface cuts into or through the principal footwall gravels. Furthermore, the Big Pebble Conglomerate occurs very close to or directly on the Kimberley Shales.

The May Reef is generally poorly developed. The small smoky quartz pebbles are characteristic. In some of the boreholes the May Reef is represented by a dark, fine-grained vitreous quartzite.

The "Chloritoid Shale" Marker is represented by arenaceous, micaceous shales, on the average about 100 feet thick. These shales contain virtually no chloritoid, but rutile is a ubiquitous secondary mineral.

The Puddingstone Reef and Big Pebble Conglomerate are generally poorly developed. In borehole K 11.47 the Big Pebble Conglomerate is represented by 7 inches of fine-grained vitreous quartzite. In this borehole the Puddingstone Reef therefore occupies the junction between the two shale zones, which are remarkably similar in this area (Plate III).

(*b*) *The Kimberley Workings:*

In the southern area near the May Reef sub-outcrop, where underground development on the May Reef is in progress, the "Chloritoid Shale" Marker is up to 180 feet thick, and may consist of alternating phases of shale, arenaceous shale, argillaceous quartzite and quartzite (borehole K 19.49); or a dominantly quartzite zone including narrow, more argillaceous layers (borehole K 24.49); or conglomeratic quartzites containing narrow layers of khaki-green, rutile-rich shale (borehole K 25.50). The geological sections of these boreholes are shown on Plate III. These shales seldom contain chloritoid.

The Big Pebble Conglomerate Zone is up to 60 feet thick. In hand-specimen the rock is indistinguishable from the lenticular conglomerates associated with the "Chloritoid Shale" Marker in this area.

The lateral change of the May Reef is very marked. In places the horizon is represented by a barren plane, developing rapidly into a robust conglomerate in which large angular chert pebbles predominate, or into a zone of grits (6 feet) in which small smoky quartz pebbles predominate, or into a well-packed medium-pebble conglomerate with the thin layer of heavy concentrate near its upper contact. The heavy concentrate in this area contains a great deal of secondary rutile, and is invariably yellowish in colour.

The May Reef is strikingly unconformable to the older beds, its distance above the Kimberley Shales varying between 50 and 250 feet. In places the erosion surface cuts through the "Chloritoid Shale" Marker into the Big Pebble Conglomerate Zone (Plate VII).

In the eastern half of the Nigel Mine the "80" Foot Marker has proved invaluable in estimating a more exact depth for the May Reef when the latter is being approached in a borehole, or in the mine workings in solving structural problems. But it is of the greatest assistance in locating the May Reef on the crests of the truncated anticlines, where the reef horizon is generally represented by a barren plane or an intermittent line of pebbles, sometimes sandwiched in between clear quartzites.

In one borehole the water supply from a pan in the vicinity was exhausted before the May Reef was reached. Calculations based on the recognition of the "80" Foot Marker made it possible for one tank of water to be transported by lorry and to prove sufficient for completing the borehole.

This conglomerate zone has some very definite characteristics in the area under review: The zone is up to 30 feet thick, in the form of a large number of inconstant grit and small-pebble bands, capped by a fairly persistent small-pebble band which lies on the average about 80 feet above the May Reef. The zone includes a white mineral in the form of small pebbles (½ inch and less), interspersed amongst quartz and chert members of the same average diameter. In other areas, e.g. on Vogelstruisbult, but more commonly in the Western East Rand, a bright green or greenish-black mineral has been observed as small pebbles in the "80" Foot Marker zone. The white mineral is very soft, with a soapy touch, and can be scratched by a finger nail. The green mineral is generally brittle, and is easily scratched by a pen-knife.

As both these rare pebbles are used as markers, their identification became of importance. The Director of the National Building Research Institute, Pretoria, kindly undertook to have these minerals subjected to an X-ray analysis. He identified the white mineral as pyrophyllite and the green one as chlorite. His reports are as follows:

(*a*) *The White Mineral:*

"The white mineral occurring in the Kimberley quartzites which was suspected to be montmorillonite was subjected to X-ray analysis and found to be pyrophyllite. Pyrophyllite has essentially the same molecular lattice structure as montmorillonite, and some varieties are known to possess considerable swelling characteristics."

(*b*) *The Green Mineral:*

"Unfortunately the large quantities of quartz in the specimen masked the peak of the other mineral to some extent, and all that could be ascertained was that the mineral belongs to the chlorite group."

The chlorite and pyrophyllite pebbles were probably derived from large crystals

in the ancient schists, the parent rocks of some of the sediments of the Witwatersrand System. It is, however, also possible for the pyrophyllite to be an alteration product of some other type of pebble, possibly chlorite. They both occur most abundantly along the same stratigraphic horizon, although seldom together in the same locality. Pyrophyllite has furthermore been observed in greatest abundance near the sub-outcrop against the Karroo rocks along the eastern and southern boundaries of the Nigel Mine.

Underground boreholes drilled through the "80" Foot Marker on Vogelstruisbult almost invariably intersected a few chlorite pebbles, some of them measuring up to ¾ inch in diameter.

One-inch pebbles of pyrophyllite have occasionally been observed in the Big Pebble Conglomerate Zone in boreholes drilled in the outcrop areas on the Nigel and Vogelstruisbult Mines.

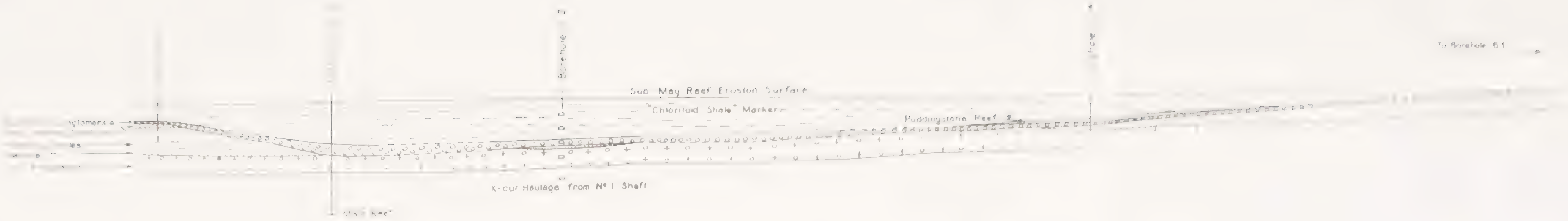
D. The Area immediately to the North of Vogelstruisbult.

This area includes Daggafontein, East Daggafontein, and the eastern half of Springs Mines.

In his discussion of Sharpe's paper (1942-1945), Bancroft (1942-1945) describes what he considers to be an infilled erosion channel immediately below the May Reef around the May Shaft on the boundary between Daggafontein and East Daggafontein Mines. He writes:

> "An exceptional feature of interest and economic importance relates to the presence of a stream channel at the base of the Kimberley Series in the vicinity of the May Shaft on the boundary between Daggafontein and East Daggafontein. At the May Shaft this ancient stream channel is apparently 600 feet to 700 feet deep and is filled with conglomerates towards its base with quartzites and shales and is capped by a Black Bar type of deposit, in the form of chloritoid-rich shales and argillaceous quartzites, that varies from 50 feet to 150 feet or more in thickness. The stream channel is so deep that in its central portion the lowest reef zone of the normal Kimberley sequence and the 120-feet thickness of quartzites with scattered chert and quartz pebbles, that in adjacent areas lies between the lowest reef zone and the top of the Kimberley Shales, and also at least most of the Kimberley shales themselves have been scoured out by the stream in the development of its valley.
>
> For some time the Black Bar type of deposit that caps this valley was considered to be the Kimberley shales and only diamond drilling and stratigraphic study have determined its true relationship.
>
> The payable 'Kimberley May' reef which is being worked on East Daggafontein and has been developed to some degree on Daggafontein has these chloritoid shales and argillaceous quartzites as its footwall for variable distances on either side of the stream channel."

In 1950 Whiteside published his paper on the geology of the Kimberley-Elsburg Series in the Brakpan-East Daggafontein area. In the western part of this area, according to his plans and sections, he regards the Upper Parent Conglomerate (U.K. 9B) as a component part of the May Reef, and he does not distinguish between the May Reef and its hangingwall leader in the eastern sector. In one place he does state that Sharpe's U.K. 9A Reef is the equivalent of the May Reef of the eastern sector and that the U.K. 9B Reef (Upper Parent Conglomerate) disappears eastwards, which agrees with the author's conclusions as set out the previous year (1949). But his map of the May Reef floor does not conform with this conclusion.



A ROUGHLY EAST-WEST SECTION ACROSS THE MIDDLE OF MARIEVALE
SHOWING THE PRE-BIG PEBBLE CONGLOMERATE SCOUR DEPRESSION

(Quartzite between Kimberley Shales and Bird Amygdaloid very thin. Compare also Borehole B.3, Plate III. Intrusions

1000 500 0 1000 2000 3000
ENGLISH FEET

Note:- For section line see plate VIII.

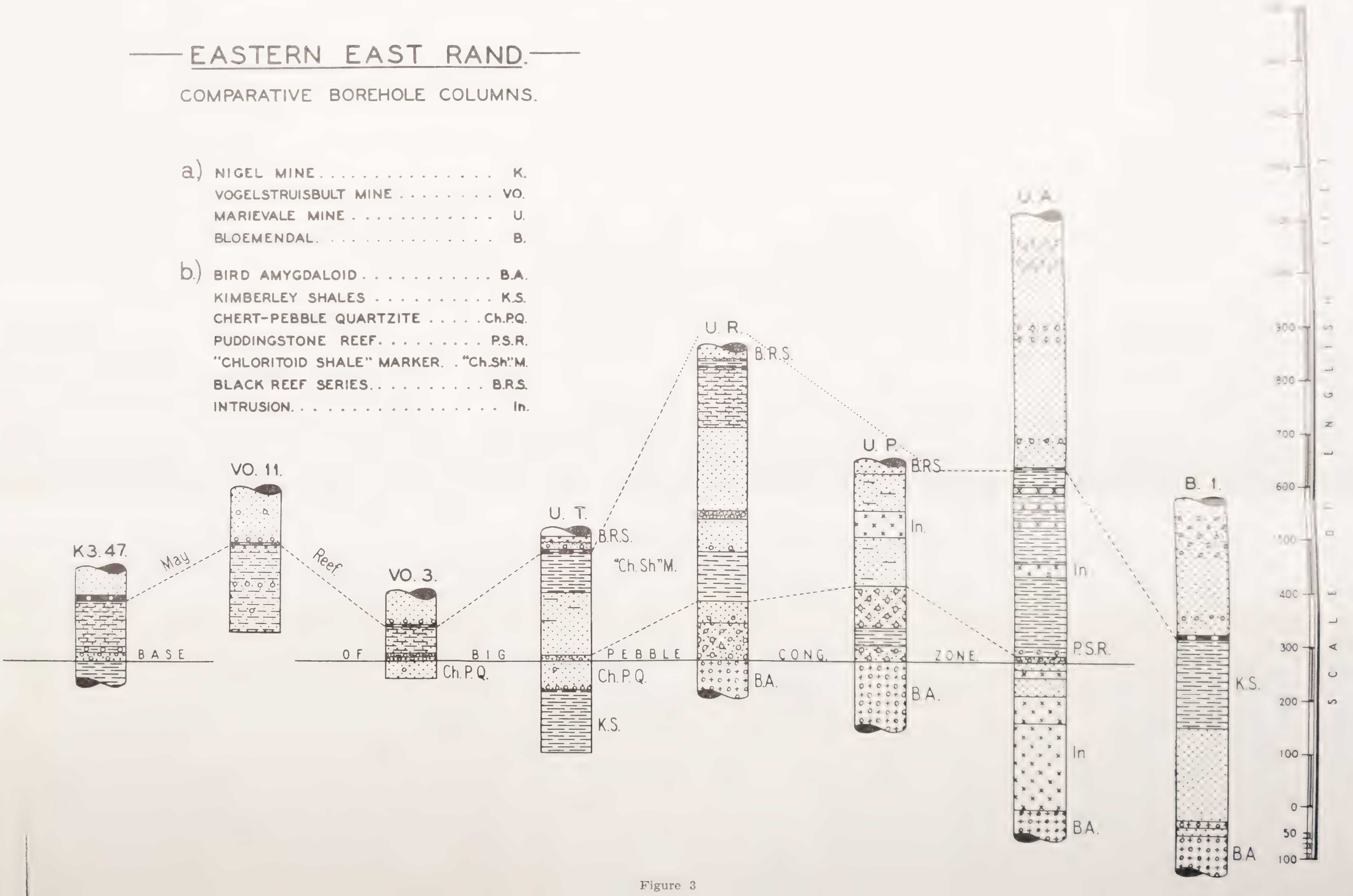


Figure 3

He gives a very interesting section through the May Shaft, previously described by Bancroft.

He interprets the M.K. 1 suite of sediments ("Chloritoid Shale" Marker) as beds which have filled erosion channels. He agrees, however, that these deposits could not everywhere be classed as the infilling of channels cut into the underlying sediments. On page 249 he writes:

> "At first it was thought that they were all deposited in channels cut into the underlying rocks but many of the smaller channel shaped deposits are better explained as remnants of truncated synclinal folds. However, there does appear to be at least one deep erosion channel, first noted in the vicinity of May Shaft, East Daggafontein."

A study of Whiteside's paper, and particularly of the very instructive plans and sections accompanying it, leads one to the conclusion that, apart from the May Shaft area, the stratigraphy of the Kimberley group of sediments in the area under review duplicates the stratigraphy of these beds on Vogelstruisbult. Important points of correspondence appear to be:

(1) Non-existence of the Upper Parent Conglomerate.
(2) Striking unconformity between the May Reef and the older Kimberley beds.
(3) The existence of an erosion surface immediately below the Big Pebble Conglomerate. The Chert-pebble Quartzite Zone is up to 200 feet thick, but in some areas the Big Pebble Conglomerate lies directly on the Kimberley Shales.
(4) The "Chloritoid Shale" Marker changes eastwards into a dominantly argillaceous zone. In the eastern half of Springs Mines the zone is largely represented by quartzites.
(5) The Puddingstone Reef appears to be of blanket character.
(6) The Big Pebble Conglomerate was again the principal contributor of gold to the May Reef.
(7) Absence of the upper quartzite phase of the Kimberley Shales.
(8) Lithologically the different phases of the May Reef correspond closely to the variations known on Vogelstruisbult.
(9) The uppermost Kimberley reef occurs on the average about 500 feet above the May Reef.

The May Shaft and Environs:

Whiteside does not show his May Reef erosion channel as an entire and separate unit on his map of the May Reef floor (Plate XLIII of his paper).

There is little doubt that the May Shaft intersected a similar, if not the same, feature as disclosed by boreholes U.R., U.A., and U.P. on Marievale (Fig. 3).

From comparison of these two areas, it appears probable that the upper sediments in the postulated May Shaft erosion channel could also be looked upon as a continuation of the "Chloritoid Shale" Marker of surrounding areas. The conglomerates towards the base of the depression are probably the Puddingstone Reef and Big Pebble Conglomerate, much thicker here than in surrounding areas, because of the depth of the floor below the general level of deposition (Whiteside, 1950, Plate XLI).

By analogy with the interpretation given to the geological sections of boreholes U.R., U.A. and U.P. on the Marievale Mine, the author therefore prefers to date the deep scouring in the May Shaft area as immediately pre-Big Pebble Conglomerate.

The scour depression on Marievale has considerable depth, but is also unusually wide. The slope of the floor on which the Big Pebble Conglomerate was deposited was thus apparently very gentle (Fig. 3).

The scour depression in the May Shaft area has steeper sides. The conglomerate (or conglomerate zone) that covers the base and sides of the May Shaft depression is noteworthy. The zone attains a maximum thickness at the deepest point of the depression (Whiteside, 1950, Plate XLI).

VII THE STRATIGRAPHY OF THE KIMBERLEY GROUP OF SEDIMENTS IN THE WESTERN HALF OF THE BASIN.

The area includes from north to south: Government Gold Mining Areas, Brakpan Mines, the S.A. Land and Exploration Company (Sallies), the western half of Springs Mines, the extreme western portion of Vogelstruisbult, Vlakfontein, Sub Nigel, the western half of the Nigel Mine, West Vlakfontein, Spaarwater, West Spaarwater, the northern portion of Witwatersrand Nigel, the Heidelberg Townlands, and portions of the farms Klippoortje 10 and Maraisdrift 4 between Nigel and Heidelberg (Plate VII).

The May Reef generally occurs from 250 to 400 feet above the Kimberley Shale Zone in the Western East Rand, except in borehole T.L. 3 on the Heidelberg Townlands, where the distance is 606 feet, and in the south-eastern portion of the Sub Nigel Mine where a minimum distance of 72 feet is known (see also Plate IV).

(*a*) The Kimberley Shale Zone has a maximum thickness of about 500 feet in this area, including the upper and lower quartzitic phases. The upper quartzite phase is known only in a few isolated places outside Brakpan Mines and Government Gold Mining Areas.

(*b*) The Chert-pebble Quartzite zone attains a maximum thickness of 400 feet in the northern portion of the area under review. The zone is up to 290 feet thick in the Heidelberg area, generally between 150 and 200 in the Sub Nigel-Spaarwater-Vlakfontein area, but in the south-eastern portion of Sub Nigel and the adjoining portion of the Nigel Mine and zone is only a few feet thick.

(*c*) The Big Pebble Conglomerate, often in the form of a zone of conglomerates, is very seldom absent from the geological column in the Western East Rand. Locally in the southern portion of the Sub Nigel Mine it has been eliminated, apparently as the result of scour before the deposition of the Puddingstone Reef, which becomes abnormally thick in such sections.

The zone is up to 160 feet thick on Government Gold Mining Areas, but thins southwards, in places represented by a few inches of fine-grained vitreous quartzite. Great individual thicknesses are, however, still met with, e.g. 122 feet in borehole T.L.3 on the Heidelberg Townlands and nearly 100 feet in some boreholes drilled in the western half of the Nigel Mine. The change in thickness often takes place rapidly. Furthermore, the individual conglomerate layers tend to change laterally into quartzite.

In the Vlakfontein-West Vlakfontein-Spaarwater area the bottom band occasionally carries payable quantities of gold.

(*d*) The Puddingstone Reef is of local occurrence in the western sector. It is fairly well developed in the Heidelberg area, where it forms a very characteristic rock on the surface. It has also been intersected in the southern portion of the Sub Nigel Mine, where its thickness fluctuates between 2 inches and 38 feet. It contains very little gold.

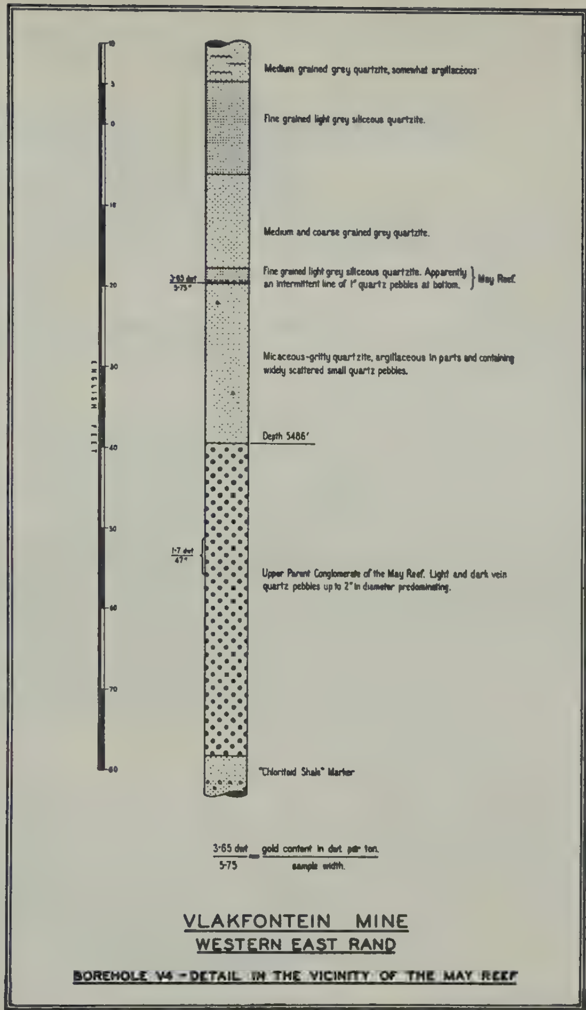


Figure 4

(*e*) *The "Chloritoid Shale" Marker:* This suite of sediments consists largely of coarse-grained quartzites and lenticular conglomerates in the Western East Rand. Locally it includes inconstant layers of shale up to 30 feet thick, and narrow hybrid conglomerates resembling the Puddingstone Reef in appearance. These shale lenses are known on Government Gold Mining Areas, have been encountered in the Vlakfontein-West Vlakfontein-Spaarwater area, but are more pronounced around Heidelberg and in the southern portion of the Sub Nigel Mine.

As a general rule it can be stated that the zone becomes finer grained and more argillaceous eastwards. This feature applies to all east-west sections across the Basin. The shales often contain chloritoid, and invariably abundant secondary rutile.

The thickness of the zone in the western sector varies between 40 and 100 feet. Locally it is only 20 feet thick, while at the other extreme, measurements up to 170 feet have been recorded.

Locally the zone does not exist, e.g. in the northern portion of Brakpan Mines, where the Upper Parent Conglomerate comes to rest directly on the Big Pebble Conglomerate.

The conglomerates associated with the zone occasionally contain traces of gold.

(*f*) *The Upper Parent Conglomerate of the May Reef:* A feature of this conglomerate zone, which attains a maximum thickness of over 50 feet, is the prevalence of white and dark-edged vein quartz pebbles up to 2½ inches in diameter. Its gold content is generally very low.

When it lies directly on the Big Pebble Conglomerate, for instance, in the northern portion of Brakpan Mines, the two horizons are distinguishable by means of the chert pebbles in the lower horizon.

An intermittently payable horizon exists along the bottom of the zone on Government Gold Mining Areas. This is Sharpe's ill-sorted U.K. 9C Reef, which he at one stage correlated with the May Reef of the eastern sector.

(*g*) The Micaceous-gritty Quartzite is up to 20 feet thick in this sector, where it forms the immediate footwall of the May Reef over large areas.

The sub-May Reef erosion surface truncates the beds below so much so that this unit and portions of the Upper Parent Conglomerate have been eliminated in some areas (Plate IV, borehole K 16. 48).

(*h*) *The May Reef:* The May Reef is seldom more than 24 inches wide, although locally widths up to 5½ feet are known. A feature of the May Reef in the western sector is the predominance of smoky and dark-edged quartz pebbles, which seldom exceed one inch in diameter. It is of economic importance on Government Gold Mining Areas, where the gold is almost invariably concentrated in the upper half of the reef, frequently in the uppermost few inches.

It rests unconformably on its footwall beds, and there is no doubt that the eroded portions of the Upper Parent Conglomerate were concentrated and spread out on the erosion surface to form the May Reef. The erosion surface apparently never cuts right through the Upper Parent Conglomerate (Plates VII and VIII, and Fig. 6).

(*i*) As mentioned before, the upper Kimberley conglomerate clusters gradually fade when traced from Government Gold Mining Areas in the north to Heidelberg in the south. In the latter area they have changed into quartzites.

A feature in the western sector is the occurrence of small chlorite pebbles in the "80" Foot Marker (Plate I).

VIII CONTINUITY OF HORIZONS IN THE EAST RAND BASIN—SUMMARY.

When studying the sediments of the Kimberley-Elsburg Series, two things become apparent:

1. The exceptional persistence of certain horizons and suites of sediments as regards both lateral extent and lithological character. The May Reef and Big Pebble Conglomerate, as well as some of the thicker zones, notably the Chert-pebble Quartzite Zone, are of very wide-spread occurrence. The Big Pebble Conglomerate is particularly constant in lithological character. These horizons have been identified to far beyond the limits of the East Rand Basin.

a) Nº 4 (CIRCULAR VENTILATION SHAFT): SECTION ALONG PORTION OF K.R.14 LEVEL WEST.

b) BOREHOLES DRILLED FROM SURFACE NEAR THE OUTCROP OR SUB-OUTCROP OF THE MAY REEF SHOWING RELATIONSHIP BETWEEN THE LATTER REEF AND THE STRATIGRAPHICALLY HIGHEST "FOOTWALL" CONGLOMERATE KNOWN.

SOUTHERN AREA OF SUB NIGEL MINE
WESTERN EAST RAND

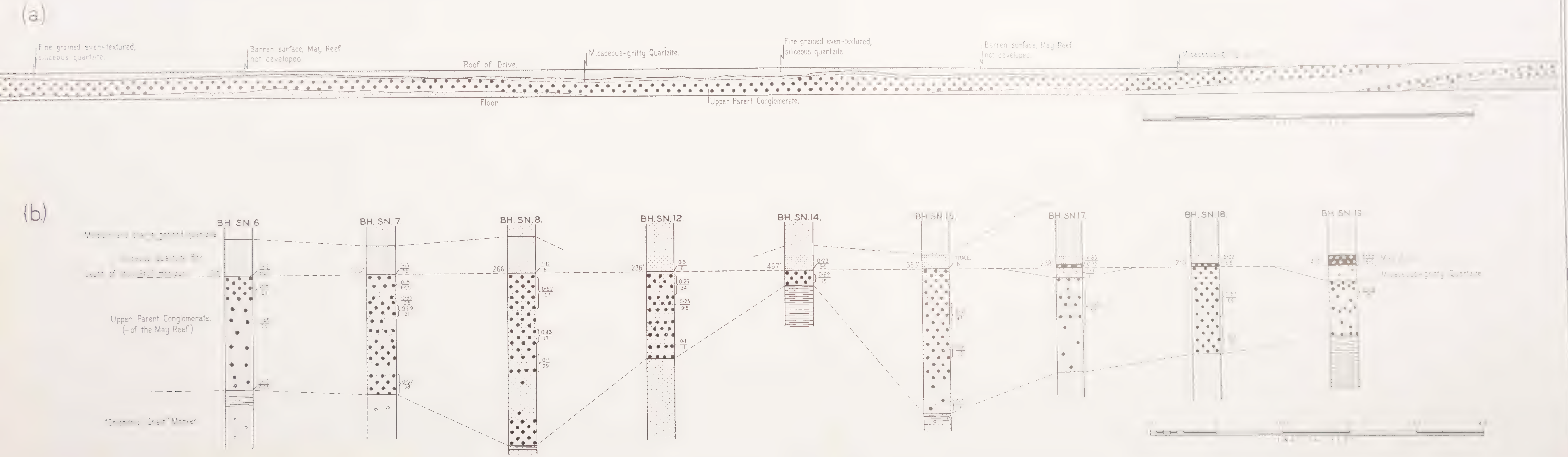


PLATE VIII

As a general rule, it appears that the absence of the thicker zones of sediments from a particular area should be attributed to removal after deposition, possibly in rare instances to non-deposition. Post-Witwatersrand erosion is not considered here.

2. In contrast to the above we have the marked lateral variation or change of facies along certain horizons, as described in the preceding pages:

(i) Individual bands of the Big Pebble Conglomerate and Upper Parent Conglomerate change laterally into quartzite locally.

Locally the Big Pebble Conglomerate Zone also includes lenses of chloritoid-rich and rutile-rich shale up to 35 feet thick. Thin lenses of shale have been observed to occur at intervals right through a 50-foot body of Big Pebble Conglomerate.

(ii) Locally the May Reef degenerates into a dark, fine-grained vitreous quartzite, notably in the eastern half of the Nigel Mine.

(iii) The suite of sediments designated the "Chloritoid Shale" Marker changes from predominantly coarse arenaceous and rudaceous sediments in the Western East Rand, to predominantly fine-grained argillaceous rocks in the eastern sector. The basal members of the zone have been deposited on a fairly even floor (Plates I and II).

(iv) The Upper Kimberley Conglomerate clusters (i.e. above the May Reef) gradually fade when traced from Government Gold Mining Areas in the north to Heidelberg in the south, in the latter area having dwindled to insignificance. The behaviour of the "80" Foot Marker is interesting, because individual pebble bands are frequently so short-lived that the entire zone could be represented by clear quartzites (Plates III and IV). Some boreholes intersected only one pebble in the zone.

IX NOTES ON THE METAMORPHISM OF THE SYSTEM, WITH PARTICULAR REFERENCE TO THE KIMBERLEY-ELSBURG SERIES.

The System has been metamorphosed regionally. The arenaceous types have been converted into quartzites and the once incoherent gravels into conglomerates, while sericitic mica and abundant chlorite have developed in the slaty types.

There are, however, two secondary minerals which deserve special attention. They are chloritoid and rutile. The latter occurs so universally along the vertical column that only the thin sections made from the pure quartzites (e.g. the Siliceous Quartzite Bar above the May Reef) failed to disclose the presence of rutile.

(*a*) *Chloritoid:*

Young (1917) comes to the conclusion that the mineral is chloritoid and not ottrelite. He writes, page 43:

> "When the chloritoid is cut approximately parallel to the base, it shows cleavages intersecting each other at about 60°. The presence of these is sufficient to distinguish the mineral from ottrelite.*"

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* It is noteworthy that two varieties of this mineral appear to be present in these sediments. Tabular and acicular crystals of the dark greenish-gray type are often twinned and frequently contain numerous inclusions. This variety which was found to contain much iron and 1·05% MnO and has a density of 3·3, may prove to be ottrelite. The other pale green and less common type probably true chloritoid—is relatively free from inclusions and occurs in coarse tabular crystals which are frequently grouped to form rosettes.

A detailed study of the distribution of this mineral in the Kimberley-Elsburg Series has been carried out. It was observed in the following horizons:

1. In the lenticular shale bodies associated with the Big Pebble Conglomerate Zone. These are up to 35 feet thick, sandwiched between the competent conglomerate layers.

2. Frequently in the matrix of the Puddingstone Reef.

3. In the "Chloritoid Shale" Marker. Chloritoid is sometimes very abundant in the argillaceous phases of this suite, not only in the narrow shale bands, but also in the thick bodies of shale. Borehole U.A. on Marievale is of particular interest. In this borehole chloritoid developed throughout 300 feet of shale. A thin section was prepared from each of the specimens collected every 10 feet.

In some sectors, notably in the eastern half of the Nigel Mine, the argillaceous phases of this suite contain virtually no chloritoid.

4. Chloritoid has been observed in the matrix of the May Reef, especially in the thin layer of heavy concentrate near the top of the reef.

5. The Argillaceous Quartzite Bar above the May Reef often contains chloritoid.

All these sediments have been folded, faulted and tilted, so that the stress conditions requisite for the formation of chloritoid could have existed. Note the compression fold in K 4. 2E Haulage (Plate V, 3 (*c*)).

Severe shearing parallel to the plane of the May Reef is evident in some areas. On East Daggafontein the shear plane locally coincides with the May Reef, crushing it, and displacing vertical structures such as dykes for distances up to 80 feet. Generally, however, the shear zone can be seen a few inches to a few feet below the reef if the latter rests on shale, and a few feet above it, if the immediate footwall happens to be a more competent rock such as hard quartzites or conglomerates.

The shear zone is particularly well-developed in the eastern half of the Nigel Mine. Chloritoid is, however, virtually absent from the "Chloritoid Shale" Marker in this area.

The mineral has not been observed in the Kimberley Shales, not even on Bloemendal 19, where the May Reef transgresses on to this argillaceous zone.

(*b*) *Rutile:*

In the East Rand Basin this mineral occurs in all the argillaceous zones and impure quartzites of the Upper Witwatersrand System, and abundantly in the Jeppestown Shales immediately below the Main Reef. It has also been observed in the conglomerates, notably in the thin heavy concentrate near the top of the May Reef.

The shale phases of the "Chloritoid Shale" Marker are particularly rich in rutile.

Numerous thin, lenticular, rutile-rich layers are associated with the Elsburg conglomerates near the top of the System.

Under the microscope the individual crystals and crystal aggregates can usually be seen only under a very high magnification. But the distribution of this mineral in the slide is always easily determined by reflected light. A high proportion of the crystals are in the form of geniculate twins. Others are fine tetragonal crystals. The crystals are frequently in the form of very slender needles, the so-called "clay-slate" needles, particularly in the Kimberley Shales.

The rutile may lie scattered through the ground-mass, or may be grouped in asteriated clusters, in bundles or sheaves, exactly as they were shed by the parent titaniferous mineral.

Leucoxene has also been observed in the sediments of the Kimberley-Elsburg Series. The mineral is easily singled out if viewed by reflected light. It occurs as irregular grains 0.2 to 0.3 mm. in diameter. It presents a whitish or light grey surface in reflected light. It is considered to be an alteration product of ilmenite. Rutile is frequently seen to "grow" on leucoxene.

Ilmenite has not been observed in the sediments of the Kimberley-Elsburg Series, but below the Main Reef, in the Jeppestown Shales, abundant ilmenite is present, usually showing an incipient alteration to leucoxene.

The three titanium minerals described above follow a very definite pattern of distribution in the Jeppestown Shales for the first 10 feet or so below the Main Reef: The shales are generally khaki-green or khaki-yellow in colour for the first few feet below the reef, gradually becoming darker, until at a distance of 8 to 10 feet below the reef they are dark greenish-grey in colour (bright green in thin section). Within this 10 feet the distribution of the three minerals under consideration is as follows:

1. The authigenic rutile crystals abound in the upper khaki-coloured zone, while an occasional whitish grain of leucoxene can be observed. Some of these grains have the characteristic yellow colour of rutile.

2. The rutile-rich zone is preceded by the darker zone in which leucoxene predominates. In this zone the rutile is frequently observed to "grow" on leucoxene.

3. Ilmenite and leucoxene occur in more or less equal proportions in the dark, greenish-grey zone at the bottom, while a little rutile may still be present. The ilmenite almost invariably shows an incipient alteration to leucoxene.

A partial chemical analysis of the khaki-shale from the Jeppestown Series gave 4 per cent TiO_2.

We are here at the junction of the competent upper Witwatersrand quartzites and the incompetent shales of the lower division. A certain amount of differential movement has taken place along this junction. It is, therefore, tentatively considered that the leucoxene and rutile developed as stress minerals from the parent detrital mineral ilmenite, rutile being the end-product. Chloritoid did not develop in these Jeppestown Shales.

As ilmenite has not been observed in the beds of the Kimberley-Elsburg Series, it would appear that the general grade of metamorphism reached in these beds is somewhat more advanced than the grade reached in the Jeppestown Series. Apparently contradicting this tentative conclusion is the fact that, regionally, the incompetent lower division suffered greater deformation than the resistant upper division.

In 1917 Young considered the bulk of the rutile which he observed in the banket to be of authigenic origin, and concluded that titaniferous iron ore appears to have been deposited along certain horizons in the Witwatersrand System, and that the rutile was afterwards derived from that mineral.

In 1939 du Toit mentioned the presence of minute rutile needles in the Kimberley Shales. He considered rutile as a secondary constituent of the banket, and concluded that it indicates the former presence of ilmenite.

H. B. Miller (1926) described "compound ilmenite-rutile" grains, the rutile being a secondary product.

He considers rutile to be a possible stress mineral in some of the world occurrences.

Harker (1939) considers the great number of minute rutile needles in cleaved slates as having been derived from the decomposition of biotite.

If, however, one considers the mode of occurrence and distribution of ilmenite, leucoxene and rutile in the sediments of the Witwatersrand System, there is little doubt

that ilmenite was the parent detrital grain, and that leucoxene and rutile are authigenic constituents formed from that mineral.

Ilmenite must have existed as an important accessory constituent in the parent rocks of the sediments of the Witwatersrand System, concentrated during the lengthy period of transportation to the site of permanent deposition. At times extensive ilmenite-sand beaches must have existed.

The occurrence of rutile leucoxene and ilmenite in certain dykes intrusive into the Witwatersrand System is interesting and has a bearing on the genesis of rutile in the metamorphosed rocks of the System: Abundant leucoxene and rutile occur in the so-called Ilmenite Diabase Dykes, which are younger than the Ventersdorp Intrusives and older than the Karroo. The Ilmenite Diabases are often very much sheared. Grains of leucoxene with unaltered ilmenite cores have been observed, but if shearing has been severe, there is usually no trace of the original accessory constituent ilmenite, but rutile abounds, occurring in bundles and sheaves and in asteriated clusters. Under the microscope these highly altered rutile-rich dykes are not always readily distinguishable from the rutile-rich metamorphosed argillaceous sediments of the Witwatersrand System.

On West Vlakfontein, borehole W.V.4 recently intersected a near vertical dyke, which contains a great abundance of fresh ilmenite. The other mineral constituents are also unaltered. This dyke is presumably of late Karroo age.

Rutile and leucoxene, although not plentiful, have also been observed in the Ventersdorp Intrusives.

X SUMMARY OF EVIDENCE FOR AN EROSION SURFACE IMMEDIATELY BELOW THE BOTTOM REEF IN THE EAST RAND BASIN.

Du Toit (1939) considers the Kimberley Shales transitional to the quartzites above and below. On Brakpan Mines and Government Gold Mining Areas one does get the impression of a gradual coarsening upwards into the Kimberley conglomerate group. But the Upper Quartzite Phase of the Kimberley Shales (a fine-grained vitreous quartzite) is generally absent away from that area. It has been found only in isolated patches in the Dunnottar-Heidelberg area, while it appears to be non-existent everywhere in the eastern sector (see Plates I, II, III and IV).

This fine-grained vitreous quartzite forms conspicuous outcrops in portions of the Greylingstad-Balfour district. No pebbles were observed in the zone, both there and in the East Rand Basin.

The coarse Chert-pebble Quartzite Zone, with the inconstant "B" Reef at the base, is always abruptly defined, irrespective of whether the Upper Quartzite Phase of the Kimberley Shales is developed or not.

The "B" Reef is strikingly unconformable to its footwall beds in parts of the Orange Free State gold field. It is not of economic importance in the East Rand Basin, but in parts of the Orange Free State gold field it is an important gold carrier. In the latter area the erosion surface sometimes cuts through the Kimberley Shales (Upper Shale Marker), eliminating also some of the Upper Bird Reefs.

While this erosion surface could not be studied in the mine workings, so that its true nature is not known, the author thinks that it satisfies Twenhofel's definition of unconformity, as quoted in Chapter IV. It appears to be a regional surface representing a rate of deposition that is zero in some areas, and less than zero in adjacent areas.

The absence of the Upper Quartzite Phase of the Kimberley Shales from certain areas is attributed to removal after deposition.

The constituent materials of the Bottom Reef in the East Rand Basin have not been derived from the floor on which it rests.

XI SUMMARY OF EVIDENCE FOR AN EROSION SURFACE IMMEDIATELY BELOW THE BIG PEBBLE CONGLOMERATE IN THE EAST RAND BASIN.

On Brakpan Mines and Government Gold Mining Areas the Big Pebble Conglomerate Zone occurs up to 400 feet above the Kimberley Shale Zone. The distance is generally less than 200 feet in the Sub Nigel-Spaarwater-Vlakfontein area, up to 290 feet in the vicinity of Heidelberg, from 3 to 30 feet in the south-eastern portion of the Sub Nigel Mine, and from 40 to 120 feet on Vogelstruisbult (see Plates III and IV).

Along the eastern boundary of the Nigel Mine the Big Pebble Conglomerate generally rests directly on Kimberley Shales, but on the Marievale side of the boundary, in borehole U.Y., the Chert-pebble Quartzite Zone is again 157 feet thick (see Plate VII for position of boreholes).

It appears, furthermore, that deep scouring took place in portions of Marievale, Daggafontein and East Daggafontein before the deposition of the Big Pebble Conglomerate Zone. This scouring resulted in the elimination of the entire Chert-pebble Quartzite Zone, the entire Kimberley Shale Zone, and even portions of the Bird Amygdaloid (Fig. 3).

The constituent materials of the Big Pebble Conglomerate were not obtained from the floor on which it rests.

XII THE UNCONFORMITY IMMEDIATELY BELOW THE MAY REEF IN THE EAST RAND BASIN.

A. The Eastern East Rand.

This erosion surface could be studied in great detail in the mine workings. It was recognised early on in this investigation because of the large number of characteristic marker horizons in the footwall of the May Reef. But the differences between certain footwall horizons are often very slight, so that if mapping is not accomplished while the exposures underground are still clean and fresh, it soon becomes an impossible task.

The unconformity is more pronounced in some sectors than in others, owing to different degrees of Pre-May Reef tectonic disturbances in different areas. Over comparatively large areas it is frequently impossible to recognise the unconformable relationship, but that does not detract from the regional conception.

(*a*) *Vogelstruisbult*

For the purpose of description, an "eroded feature" would imply an area in which the sub-May Reef erosion surface has cut into or through the Puddingstone Reef and Big Pebble Conglomerate.

1. *The Area East of No.* 1 *Shaft.*

The relationship here is one of an erosion surface truncating well-defined anticlines and synclines, which generally trend in a direction N.30°W-S.30°E. But at least one large eroded feature trends in a NE-SW direction (Plates VI and VII).

The angular difference between the May Reef and the older gravels is usually small, although locally, measurements up to 47° were recorded. These large angles do not persist in depth, however, as clearly illustrated by the geological section along K 5.5 Haulage (Plate V, 2).

Locally the May Reef footwall sediments were folded very intricately, in one case giving rise to a complex, now truncated,tdome (Plate VI, K 4.1B Drive).

It is not possible to calculate or estimate the amplitudes of these eroded anticlines to any degree of accuracy, as the folding appears to have been accompanied by a general but extremely gradual uplift of the entire depositional floor. The structurally high areas were planed down more or less flush with the adjacent structurally low areas. The coarse and heavy fractions of the detritus thus produced (including the gold and other heavy minerals shed by the older gravels) were deposited on the flanks of the truncated anticlines, but largely on the sub-May Reef synclines (Plates V, 1 and 2, and VI). Occasionally a thin veneer of "lag gravel" found permanent deposition on the crests of the eroded anticlines (Plate V, 4 (*a*) and (*b*)).

It therefore appears that from a certain stage in the process the floor coinciding with the sub-May Reef synclines stood at a slightly lower elevation than the eroded crests of the anticlines, allowing a thickness of May Reef equal to the distance between the wave base and floor. This distance varies between a fraction of an inch and 10 feet.

Some of these eroded features and the distribution of the gold in the May Reef relative to them are shown on Plate VI.

The following conglomerate horizons contributed to the mineral content of the May Reef in this area:

(i) Lenticular conglomerates associated with the "Chloritoid Shale" Marker.
(ii) The Puddingstone Reef.
(iii) The Big Pebble Conglomerate, which was the principal contributor of gold.
(iv) Lenticular conglomerates in the Chert-pebble Quartzite Zone.

It is not known whether the Upper Parent Conglomerate of the western sector ever existed in the area under review (Plate I).

Notes on Individual Eroded Anticlines—Plate VI.

Eroded Anticline No. 1

Erosion apparently did not penetrate the Big Pebble Conglomerate. The Puddingstone Reef, therefore, appears to have been the principal source gravel in this area, but because of its generally low gold content, a payable May Reef could not have formed over extensive areas. The small payable areas that are arranged on the north-eastern limb of this truncated anticline are considered to be related to the Puddingstone Reef.

Eroded Anticline No. 2

Geological sections measured across portions of this eroded feature are shown on Plates V, 1 and 4.

The May Reef has a low gold content immediately on and along the south-western limb of this feature. The reef is, however, more than two feet wide in places, and locally includes a high proportion of large chert pebbles, for instance, near the

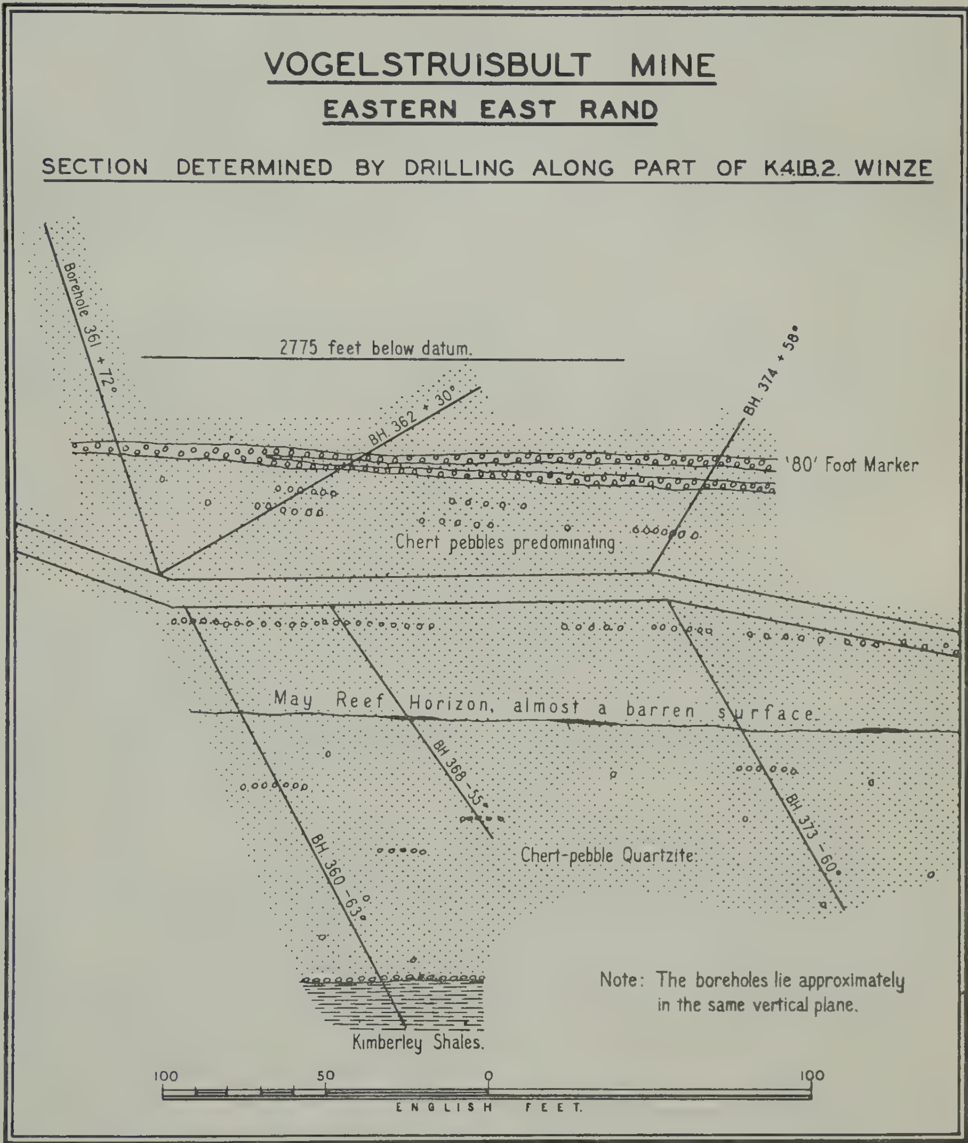

VOGELSTRUISBULT MINE
EASTERN EAST RAND
SECTION DETERMINED BY DRILLING ALONG PART OF K4.IB.2. WINZE
Borehole 361 + 72°
2775 feet below datum.
BH. 362 + 30°
BH. 374 + 58°
'80' Foot Marker
Chert pebbles predominating
May Reef Horizon, almost a barren surface.
BH 368 -55°
BH 373 -60°
BH 360 -63°
Chert-pebble Quartzite.
Note: The boreholes lie approximately in the same vertical plane.
Kimberley Shales.
100
50
0
100
ENGLISH FEET.

Figure 5

start of K 4.2B Haulage (Plate V, 1 (*b*)). These were obtained from the Big Pebble Conglomerate in the immediate vicinity.

As a general rule very little May Reef was deposited on the flat crest of this truncated anticline. It is often very difficult to follow this barren plane successfully, particularly if a well-developed May Reef Hangingwall Leader occurs immediately above the Big Pebble Conglomerate. If the former also includes a few large chert pebbles, as sometimes happens, the May Reef horizon cannot be singled out.

The complex eroded dome at the north-western extremity of the feature under review is covered by a thin veneer of May Reef, very rich in gold and containing a great amount of soft carbon. For distances of 3 feet or more the reef does not include a single pebble, but is in the form of a thin veneer of sulphides, carbon, quartz grains, and contains the heavy minerals gold, zircon, rutile, chromite, etc. The crests of the eroded anticlines, although generally areas of non-deposition, are therefore never eliminated beforehand as being of no interest.

The richest May Reef which appears to be related to the No. 2 eroded anticline occurs on and along its north-eastern limb, and extends for some 600 feet on to the associated syncline (Plate VI (*b*)). The highest gold values are generally confined to the uppermost few inches (Plate V, 4 (*c*)), and are associated with a thin layer of fine-grained heavy minerals, which exists in sedimentary continuity with the thin veneer of rich May Reef on the eroded dome.

Eroded Anticline No. 3

Along K 4.3 Haulage the pay shoots are confined to the sub-May Reef synclines. The richest May Reef existing in the mine has been found here.

Eroded Anticline No. 4 (*North of the tear fault*)

There was considerable difficulty in locating the May Reef horizon in this area. Two years elapsed before it was discovered that the areas to the north and south of the fault were not in sedimentary continuity (Plate VI). The geological section of K 4.1.B.2 Winze (Fig. 5) illustrates the problem encountered here. Boreholes were drilled upwards and downwards from this winze in search of a known marker horizon. A tentative interpretation was given at the completion of boreholes 373 and 374. The end was then directed downwards and eventually intersected the May Reef horizon on the Big Pebble Conglomerate near the common boundary with East Daggafontein Mine.

It was thus realized that K4.1.B.2 Winze had entered on to an eroded feature of comparatively large amplitude. K.5 Haulage (Plate V, 3 (*a*)) thereafter experienced no difficulty, and eventually K5.5 Haulage (Plate V, 2) exposed payable May Reef on the easily recognisable chloritoid shale. More than half a million tons of ore reserve have subsequently been blocked out westwards from this eroded feature.

We have thus in this area striking examples of the provenance of the May Reef, i.e. stratigraphically older auriferous gravels. Owing to the small difference in angle of deposition between the May Reef and the older gravels, and the consequent sheet erosion effect, an enormous supply of gravelly material was made available for concentration.

Several more eroded features have been exposed in the mine workings south-east and South West of No. 4 Shaft. The general rule also applies that the structurally high areas were areas of non-deposition as far as the May Reef is concerned, except in one comparatively large area south-west of No. 4 Shaft, where a wide robust May Reef found permanent deposition on the crest of an eroded anticline (Plate X,

Ph. 6). In such instances the May Reef is distinguished from the Big Pebble Conglomerate by its higher gold content, and visually by the fact that chert pebbles may be more prevalent in the lower horizon. The condition described above could be explained in one of two ways: (i) either the material derived and transported from this terrain was returned or (ii) was derived from another eroded area in the vicinity.

The distribution of chert pebbles in the May Reef follows a definite pattern in the area east of Vogelstruisbult No. 1 Shaft: on and near the eroded features chert pebbles and cobbles are common, at times predominating over the quartz members (Plate X, Ph. 2 and 8). With increasing distance from the eroded areas the chert pebbles become inconspicuous, i.e. smaller and often absent (Plate X, Ph. 1 and 7). The chert pebbles in the footwall gravels are generally large, so that their disappearance from the May Reef in a given direction from an eroded feature could be the result of sorting of the pebbles according to size.

2. *No.* 2 *Shaft and the Outcrop Area*

No eroded features have been discovered south-westwards towards No. 2 Shaft or in the outcrop area. The exceptional variation in the thickness of the "Chloritoid Shale" Marker in the outcrop area is attributed in part to the unconformity under review.

Eroded features might have existed behind the line of outcrop or sub-outcrop of the May Reef, subsequently obliterated by pre-Transvaal or pre-Karroo erosion (Plate VII).

(*b*) *Marievale Mine*

In the Draaikraal sector adjoining the Nigel Mine, borehole U.Y. intersected an eroded anticline of comparatively large amplitude (Plate VII).

Drilling on Bloemendal showed the May Reef to lie on Kimberley Shales. It is therefore possible that the May Reef also lies on Kimberley Shales in the adjoining portion of Marievale (Plate VII).

Elsewhere on Marievale the May Reef rests on the chloritoid-bearing shales over large areas. Note the thickness of the "Chloritoid Shale" Marker in boreholes U.R., U.P., U.T. and U.A. (Fig. 3).

(*c*) *The Nigel Mine*

In the southern portion of the mine, where underground development is in progress, the May Reef is strikingly unconformable to its footwall beds. The robust lenticular conglomerates associated with the "Chloritoid Shale" Marker are considered to have been the principal parent gravels in this area, although locally the erosion surface cuts into the Big Pebble Conglomerate Zone. Seeing that we are here very near the line of demarcation between the western and eastern sectors of the Basin, the Upper Parent Conglomerate of the western sector might have been an important contributor to the mineral content of the May Reef in the area under review (Plate VII).

(*d*) *The Area to the North of Vogelstruisbult*

This area includes Daggafontein, East Daggafontein and the eastern half of Springs Mines.

The sub-May-Reef erosion surface has cut into and through the Puddingstone Reef and Big Pebble Conglomerate over fairly large areas in this sector. There can be no doubt that these two horizons were the principal parent gravels, the Big Pebble Conglomerate having been the most important contributor of gold to the May Reef. It is not possible to relate a particular payable area on the May Reef to a specific

eroded feature. The coarse and heavy fractions of the material obtained from the footwall rocks were deposited along narrow well-defined zones, trending in a north-west south-east direction (Whiteside, 1950, Plates XLIII and XLV).

B. The Western East Rand

In the eastern sector the "Chloritoid Shale" Marker is the stratigraphically highest footwall series below the May Reef (Plate II).

In the western sector the Upper Parent Conglomerate and the Micaceous-gritty Quartzite intervene between the "Chloritoid Shale" Marker and the May Reef (Plate I). The May Reef has the characteristic Micaceous-gritty Quartzite as its immediate footwall over large areas in this sector. In places the reef transgresses on to the Upper Parent Conglomerate. The unconformable relationship is thus established (Plate VII).

The Upper Parent Conglomerate disappears eastwards along the dividing line between the two sectors. It is not known to what extent it existed in the eastern sector.

The relationship between the May Reef and its floor in the western sector is one of a basal conglomerate blanketing an erosion surface that truncates shallow flexures. The angular difference is therefore small, amounting at the most to a few degrees. On the structurally high areas the erosion surface cuts well into, but apparently never right through the Upper Parent Conglomerate. The second-generation pebbles of the May Reef have suffered a reduction in diameter of more than 50 per cent.

"Windows" of the Upper Parent Conglomerate have been located in the following areas:

(*a*) *Around Sub Nigel No.* 3 *Shaft*

A drive on May Reef intersected two large eroded features. The erosion surface cuts well into, but never through the Upper Parent Conglomerate (Plate VII and Fig. 6).

(*b*) An extensive eroded area exists in the southern portion of the Sub Nigel Mine, extending also into the Nigel Mine, as indicated by boreholes S.N.6 to S.N.19 (Plates VII and VIII (*b*)). The geological sections of these boreholes indicate the following sequence of events that led to the formation of the May Reef:

1. Uplift of the depositional floor and warping of parts of it, to above a base level of deposition.

2. Concomitant erosion extending to near that level, which resulted in the concentration of the gravelly material obtained from the slightly auriferous Upper Parent Conglomerate.

The second-generation pebbles suffered a reduction in diameter of more than 50 per cent.

3. Permanent deposition of the concentrate in an adjacent slightly lower-lying area.

Note the gold content of the May Reef in boreholes S.N.17, S.N.18 and S.N.19 (Plate VIII (*b*)).

(*c*) Around the C.V. Shaft of the Sub Nigel Mine the May Reef horizon is represented by a barren plane, sometimes on the Micaceous-gritty Quartzite and sometimes on the Upper Parent Conglomerate (Plates VII and VIII (*a*)).

(*d*) Along the outcrops between Nigel and Heidelberg, the floor of the May Reef could be studied at intervals.

An extensive eroded feature was recognised in the southern portion of Spaarwater Mine. The May Reef is apparently not developed in this area.

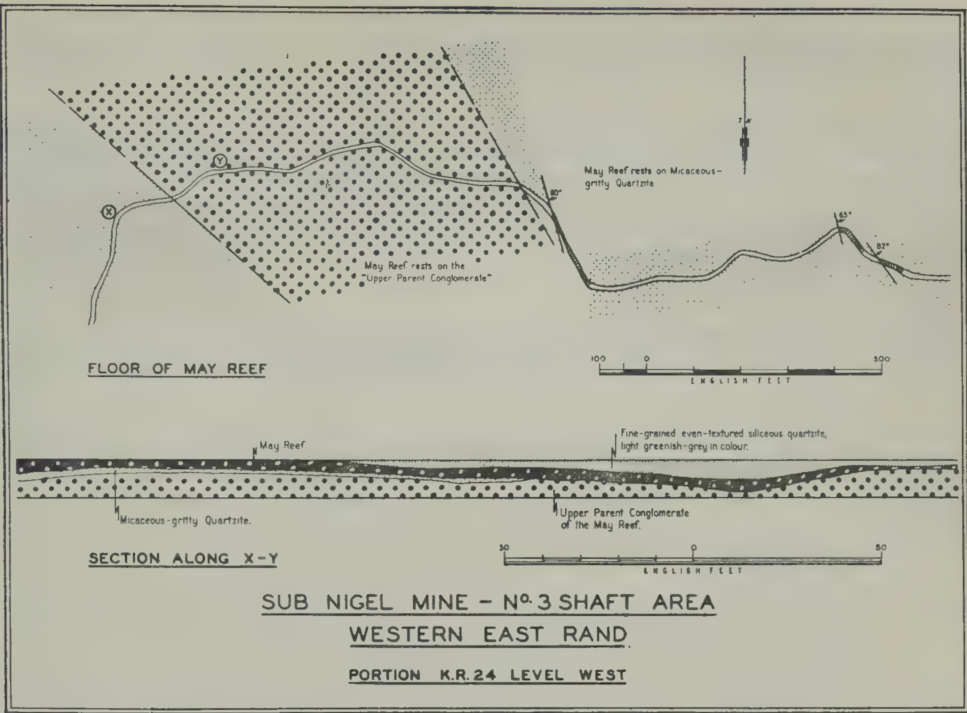


Figure 6

Further south, on Maraisdrift 4, the May Reef is again well developed, on the Micaceous-gritty Quartzite as footwall (Plate VII).

(*e*) Most of the deep boreholes drilled in the Heidelberg-Dunnottar area showed the May Reef lying on the Micaceous-gritty Quartzite. Borehole T.L.3 on the Heidelberg Townlands intersected an eroded feature. The May Reef is not developed in this borehole.

(*f*) A study of the literature strongly suggests that in the area to the north of Vlakfontein (Sallies to Government Gold Mining Areas), the May Reef owes its existence to the same sequence of events as described above.

Note: The author believes that many more eroded features exist in the Western East Rand.

XIII THE KIMBERLEY-ELSBURG SERIES IN OTHER AREAS, WITH SPECIAL REFERENCE TO THE KIMBERLEY GROUP OF SEDIMENTS.

A. Greylingstad-Balfour District, i.e. the Area to the South of the Sugarbush Fault (Fig. 1).

The area under review stretches from Modderfontein 56 in the west to the old Heidelberg-Roodepoort Mine in the east. This mine is also known as East Nigel Gold Areas.

For descriptive purposes this outcrop area is divided into three parts, a western, a central and an eastern sector. Geologically the area could probably best be divided into two sectors, as in the case of the East Rand Basin.

(*a*) *The Western Sector*

This sector includes the outcrop area on Modderfontein 56, Malanskraal 73, Driefontein 280, and Tweefontein 98 (geological map by Rogers, 1922).

This area has not been examined, so that nothing concrete about it is known, except that exposures are good, and that there are numerous prospect trenches and pits, particularly on Malanskraal.

(*b*) *The Central Sector*

This sector includes the outcrops on Rietfontein 244 and Daspoort 120. The old Balfour Gold Mine is situated on the latter farm (geological map by Rogers, 1922).

1. *Rietfontein* 244

The Kimberley reef outcrops are located about 3 miles south-west of Balfour railway station. The following stratigraphic units of the Kimberley group of sediments have been recognised, striking east-west and dipping to the north at fairly steep angles:

(i) The Kimberley Shales proper, immediately to the south of a prominent quartzite ridge.

(ii) The quartzite ridge mentioned above is formed by the Upper Quartzite Phase of the Kimberley Shale Zone. These quartzites are also known on Brakpan Mines and Government Gold Mining Areas.

(iii) The Chert-pebble Quartzite Zone. These beds do not form good outcrops, but positive proof was obtained from an old excavation immediately to the south of the railway line. From the rock dumps at the sites of the inclined shafts immediately to the north of the railway line fresh specimens were obtained.

(iv) The Big Pebble Conglomerate. This robust conglomerate is composed almost entirely of large discoidal and angular chert pebbles, exhibiting a complete lack of imbrication. It is very similar in appearance to the intersection of the Big Pebble Conglomerate by borehole E.9H on the Far West Rand.

(v) The Puddingstone Reef occurs a short distance above the Big Pebble Conglomerate and forms a very good outcrop. In the weathered state the rock is reddish-brown in appearance, studded with the various coloured pebbles. Outcrops of the Puddingstone Reef were first studied on the Heidelberg Townlands.

(vi) The May Reef, which is a small-pebble reef up to 12 inches wide, rests on the Puddingstone Reef.

2. *Daspoort* 120—*the Old Balfour Gold Mine*

Judging by the rock dumps at the shaft site it would appear that the bulk of the underground work had been carried out on the May Reef.

Very fine outcrops could be studied, and a few old winzes on the May Reef were accessible for short distances.

The following horizons were recognised:

(i) The Chert-pebble Quartzite Zone.

THE KIMBERLEY-ELSBURG SERIES

CORRELATION TABLE OF INDIVIDUAL STRATIGRAPHIC UNITS WITH SPECIAL REFERENCE TO THE KIMBERLEY ECONOMIC ZONE

THE LEFT-HAND VERTICAL COLUMN SUGGESTS A SYSTEM OF NOMENCLATURE FOR GENERAL USE OVER THE ENTIRE STRUCTURAL BASIN. OTHER LOCAL NAMES BEING USED IN THE VARIOUS SECTORS ARE GIVEN IN THE RELEVANT COLUMNS. EARLIER ERRONEOUS CORRELATION IS THUS ILLUSTRATED IN SOME CASES.

	SYSTEM OF NOMENCLATURE SUGGESTED FOR GENERAL USE ↓	EAST RAND BASIN.		GREYLINGSTAD – BALFOUR DISTRICT. (i.e. Area south of the Sugarbush Fault)			WEST RAND AREA. (i.e. Area to the north west of Witpoortje Fault)			WEST WITWATERSRAND AREAS.	KLERKSDORP AREA	ORANGE FREE STATE		VREDEFORT AREA
		Western Sector	Eastern Sector	Western Sector	Central Sector	Eastern Sector	Luipaardsvlei Estates	West Rand Consolidated	Randfontein Estates	[illegible]		South-east of Odendaalsrus.	South of Odendaalsrus.	
	Elsburg Group of Conglomerates.	+	+								Upper Gold Estates Series?	"V.S.1 Agglomerate = Conglomerate"?	"Ventersdorp Basal Conglomerate"	
	Kimberley-Elsburg Intermediate Quartzites.	+	+	+	+	+	+	+	+	+	Middle Gold Estates Series?			
KIMBERLEY CONGLOMERATE GROUP – UPPER	Conglomerate Zone.	+ U.K.3 Conglomerate Zone.	+ U.K.3 Conglomerate Zone.						+ "Upper Horsham Reefs"	+				
	Quartzite.	+ U.K.4 Quartzite Zone.	+ U.K.4 Quartzite Zone.							+				
	Conglomerate Zone.	+ U.K.5 Conglomerate Zone.	+ U.K.5 Conglomerate Zone.							+				
	Quartzite.	+ U.K.6 Quartzite Zone.	+ U.K.6 Quartzite Zone.						+	+				
	'80' Foot Marker.	+ '80' Foot Leader. U.K.7 Conglomerate Zone.	+ '80' Foot Leader. U.K.7 Conglomerate Zone.				+	+	+ One of the "Upper Horsham Reefs."	+		+ "V.S.5 Agglomerate = Conglomerate"		
	Coarse-grained Quartzite.	+ U.K.8 Quartzite Zone.	+ U.K.8 Quartzite Zone.		+	+	+	+	+	+		+		
	Argillaceous Quartzite Bar.	–	+							–				
	May Reef Hangingwall Leader.	–	+ Immediate Hanging-wall Leader.				+	+	+	–				
	Local discontinuity resulting in the elimination of the May Reef in some small areas.													
	Siliceous Quartzite Bar.	+ U.K.9A Marker.	+			+				–				
	May Reef.	+ U.K.9A Reef.	+		–	+	+ Battery Reef	+ Battery Pay Band	+ Horsham Reef.	+	+	+ "[illegible] Reef"	+ "[illegible]"	
	Regional unconformity. In some areas the May Reef even transgresses on to the Kimberley Shales, representing the cumulative effect of more than one stratigraphic break. It appears that in the O.F.S. the May Reef may even rest on quartzites which normally lie ± 1000 feet below the Basal Reef; More work is however necessary to establish this relationship for certain. "Eroded footwall material" concentrated to form the May Reef.													
KIMBERLEY CONGLOMERATE GROUP – MIDDLE	Micaceous gritty Quartzite.	+	–		–	–				–				
	Upper Parent Conglomerate (– of the May Reef).	+ U.K.9B Reef. Lower May Reef.	–		–	–								
	Ill-sorted Reef.	+ U.K.9C Reef.	–		–	–				–				
	Local discontinuity resulting in the concentration of "eroded footwall material" to form an economic but ill-sorted reef (U.K.9C) on Government Gold Mining Areas.													
	"Chloritoid Shale" Marker	+ Upper M.K.1 Sediments.	+ Upper M.K.1 Sediments.		+	–		+	+	+		+ "Chloritoid Shale" plus "E.C. Conglomerate Zone"		
	Local discontinuity resulting in the elimination of the Puddingstone Reef and Big Pebble Conglomerate.													
	Puddingstone Reef.	+ Bastard Reef (of local occurrence).	+ Scattered Pebble Reef.		+	–				–		+ (apparently of local occurrence)		
	Local discontinuity resulting in the elimination of the Big Pebble Conglomerate.													
	Siliceous Quartzites and/or Argillaceous Quartzites and/or Small pebble bands.	+	+		+	–				–				
	Big Pebble Conglomerate.	+ Large Pebble Reef or M.K.2 Conglomerate.	+ Large Pebble Reef or M.K.2 Conglomerate		+	–	+ Boulder Reef.	+ Boulder Reef.	+ Lindum Reef.		[illegible]	[illegible]	[illegible]	
	Regional Unconformity. Big Pebble Conglomerate may occur a few hundred feet above the Kimberley Shales, or may lie directly on the shales; the Kimberley Shales may also be completely eliminated. As compared to the May Reef, the Big Pebble Conglomerate consists of "foreign" material.													
KIMBERLEY CONGLOMERATE GROUP – LOWER	Chert-pebble Quartzite Zone.	+ M.K.3 Quartzite. Footwall (cherty) Quartzite.	+ M.K.3 Quartzite. Footwall (cherty) Quartzite.		+	–	+	+				[illegible]	[illegible]	
	Bottom Reef (Inconstant).	+	+		+	–	+			[illegible]		[illegible]	[illegible]	
	Regional unconformity. Erosion surface cuts through the Kimberley Shales in places. In the O.F.S. some [illegible] content of the Bottom Reef.													
UPPER MAIN BIRD SERIES	Upper Transition Phase of the Kimberley Shales.	+ L.K.1	–		+	–								
	Kimberley Shales Proper.	+ L.K.2	+ L.K.2	+	+	+	+	+	+	+		[illegible]	[illegible]	
	Lower Transition Phase of the Kimberley Shales.	+ L.K.3	+ L.K.3											
	Upper Main-Bird Quartzite.	+	+	+	+	+	+	+	+	+	+	+		

REMARKS-

WESTERN EAST RAND / EASTERN EAST RAND } The dividing line runs through the middle of Nigel Mine, immediately east of Vogelstruisbult No. 3 Shaft, [illegible] Springs Mines in a north-westerly direction, and then through a point somewhere east of Government G.M. Areas.

GREYLINGSTAD - BALFOUR {
- WESTERN SECTOR- - - outcrop area on Modderfontein 56, [illegible]
- CENTRAL SECTOR- - - outcrop area on Rietfontein 244 and [illegible]
- EASTERN SECTOR- - - East Nigel G.M. Ltd., East Wits G.M. Ltd., Mex River G.M. [illegible]

\+ indicates positive identification of a particular stratigraphic unit. Insufficient evidence indicated by a [illegible]

– indicates apparent entire absence of a particular stratigraphic unit from the relevant sector, or absence from a [illegible]

(ii) The Big Pebble Conglomerate.

(iii) The Puddingstone Reef. Very typical specimens were obtained from the rock dumps.

(iv) A fine-grained, khaki-green, rutile-rich shale, which is a phase of the "Chloritoid Shale" Marker. A similar rock occurs in the same stratigraphic position in borehole E.9H on the Far West Rand.

(v) The May Reef which rests alternately on the khaki shales and the Puddingstone Reef. The May Reef is up to 3 feet thick and includes fairly wide bands of a dark, vitreous quartzite. The majority of the pebbles are dark, smoky quartz.

The May Reef on Daspoort 120 is similar in all respects to the May Reef intersected by boreholes V°.3 to V°.13 in the outcrop area on Vogelstruisbult Mine. The former area lies due south of the latter.

(*c*) *The Eastern Sector*

This sector covers the area to the south and south-west of Greylingstad up to a distance of 13 miles from the village (Fig. 1).

It includes:

1. East Nigel Gold Areas (The old Heidelberg-Roodepoort Mine).
2. East Wits Gold Mining Areas, on the farm Rooiwal 295.
3. New Rand Reefs, and South-East Wits Gold Mining Company.

A portion of this area was recently worked under the name of Hex River Gold Areas.

1. *East Nigel Gold Areas*

The outcrops are located 2½ miles due south of Greylingstad. The beds strike roughly north-south, dipping to the west at angles varying between 20° and 35°.

The reef on the Kimberley Shales is beyond any doubt the May Reef. The identification of the Kimberley Shales is confirmed by outcrops of the Bird Amygdaloid in the vicinity.

The May Reef is up to 3 feet wide. Most of the pebbles are of the dark, smoky quartz variety, with very occasional chert members. They do not exceed one inch in diameter, and are generally much smaller. The reef body includes fairly wide bands of dark, fine-grained, vitreous quartzite. Texturally the reef compares well with the May Reef in certain parts of the East Rand Basin, notably on Marievale, portions of Vogelstruisbult and East Daggafontein, and the eastern half of the Nigel Mine.

2. *East Wits Gold Mining Areas on the farm Rooiwal* 295

The outcrops occur due south-west of Greylingstad, at a distance of about 13 miles from the village. The beds strike in a direction N22°W—S22°E, and dip at very steep angles to the north-east.

The reef on the Kimberley Shales is undoubtedly the May Reef.

A second shale horizon, which appears to be the Jeppestown Shale, outcrops further updip.

The Bird Amygdaloid was not identified.

A number of winzes were sunk on the reef to depths of about 50 feet, and about 1,000 feet of lateral driving accomplished.

The quartzites above the reef horizon leading up to the Ventersdorp Lava compare well with the May Reef hangingwall as known in the Heidelberg area.

The May Reef has a very constant thickness of about 3 feet. It is a well-packed conglomerate bed, with a minimum of internal "waste". The bulk of the pebbles are of smoky quartz, seldom exceeding one inch in diameter. The occasional chert members are easily overlooked in the dark matrix of the reef.

3. *Hex River Gold Areas* (*Fig.* 1).

The outcrops are located 12 miles from Greylingstad, in a direction slightly west of south. An old winze near the Greylingstad-Villiers road was examined. A poorly developed reef, 3 feet wide, lies on a weathered shale. Correlation with the May Reef is suggested by a fine-grained, vitreous quartzite occurring immediately above the reef. It is responsible for a low, but well-defined ridge on the surface. Considering the neighbouring areas described above, there remains but little doubt that we are dealing with the May Reef, again lying on Kimberley Shales.

It is interesting to note that, geologically, the eastern portion of the Greylign-stad-Balfour sector corresponds very closely to the farm Bloemendal 19 and the adjoining portion of the Marievale Mine north of the Sugarbush Fault.

B. Rose Deep Mine to East Rand Proprietary Mines (between Germiston and Brakpan).

No information about the Kimberley-Elsburg Series could be obtained for this section of the Witwatersrand gold field.

C. The Simmer and Jack Mines to the Witpoortje Break.

This section embraces most of the Central Rand proper.

G. Carleton Jones (1936) writes about this sector:

> "The base of the Kimberley Reef zone is variously given as from 3,200 to 4,000 feet or more above the Main Reef. The group has not been properly delimited in this area and the reef zone is variously reported as being anywhere from 520 to 800 feet in thickness and to include as many as 13 or even 17 individual reefs, with widths ranging from 18 inches to 11 feet, and with pebbles up to six inches and more in size and frequently accompanied by much pyrite. These reefs were worked in the past in the old Great Britain Mine and in the Consolidated Main Reef. A certain amount of development work, mostly on the surface, has been done on the Kimberley Reef in other mines, but results have not been encouraging."

The South Deep Shaft on the Simmer and Jack Mines was recently examined. But owing to the awkward working conditions in the shaft, its age and the consequent condition of the walls, the examination was not very satisfactory. Nevertheless, some positive information was obtained (Plate IX):

1. The Kimberley Shales are transitional upward into quartzite. To all intents and purposes there could be a gradual coarsening upward into the Kimberley conglomerate group.

2. A 5-foot band of fine-grained, khaki-green, rutile-rich shale, also containing chloritoid, occurs at a depth of 1,111 feet.

3. At 945 feet the reef is tentatively considered to be the May Reef. It gave the highest individual value of the Kimberley reefs intersected by the shaft. The reef is 90 inches wide, with the highest values limited to the upper band. It is a well-packed quartz-pebble conglomerate, the bulk of the pebbles being up to 1½ inches in diameter.

4. Above the 945-foot reef the sediments could be resolved into quartzite zones and conglomerate clusters, corresponding to Sharpe's sub-divisions on Government Gold Mining Areas. (A borehole from surface just completed has confirmed the above analysis.)

D. The Far West Rand

Strictly speaking the West Rand should now be considered, but because a solution was first found for the Far West Rand, it will be discussed first. The correlation in the West Rand area is based to a great extent on the geological section of borehole E.9H near Blyvooruitzicht Mine. An examination of this borehole, situated on the farm Kleinfontein No. 36, showed that the Kimberley group of sediments on the Far West Rand could be resolved into the same sub-divisions as in the East Rand Basin (Plate IX). That establishes correlation.

The individual units that were recognised are as follows:

1. The Kimberley Shale Zone with its upper and lower quartzite phases between depths of 4,990 and 5,594 feet, a total borehole thickness of 604 feet.
2. The horizon of the Bottom Reef at 4,990 feet.
3. The Chert-pebble Quartzite Zone between depths of 4,761 and 4,990 feet.
4. The Big Pebble Conglomerate (3 feet wide) at 4,761 feet.
5. The "Chloritoid Shale" Marker between depths of 4,603 and 4,758 feet, a total borehole thickness of 155 feet.

Lithologically the suite varies rapidly along the vertical column, and most probably also laterally. Many of the contacts between individual units are sharp. The different phases which could be recognised are:

(i) Fine-grained khaki-green rutile-rich shale between depths of 4,650 and 4,660 feet.
(ii) Narrow chert-pebble bands, probably lenticular.
(iii) Widely scattered chert pebbles (½ inch) in a greenish somewhat argillaceous quartzite.
(iv) Concentrations of black chert fragments in quartzite.
(v) Coarse-grained khaki-green quartzite, with matrix consisting mostly of chlorite, rutile and a little sericite.
(vi) Numerous narrow layers of fine-grained, whitish, vitreous quartzite, having sharp contacts. Under the microscope it is seen to be a nearly pure quartzite, the quartz grains forming a very close mosaic.

6. The May Reef occurs at a depth of 4,603 feet.
7. The "80" Foot Marker is represented by a 6-inch pyritic pebbly grit at 4,487 feet.

E. The West Rand Area

Luipaardsvlei Estates, West Rand Consolidated Mines and Randfontein Estates are discussed under this heading (Jones, 1936, Plan B).

Facts and ideas were obtained from the Presidential address to the Geological Society of South Africa by G. Carleton Jones (1936). Additional information was obtained from publications by C. S. McLean (1931-1936) and C. W. Pegg (1950).

(*a*) *Luipaardsvlei Estates*

Considering the geological section of borehole E.9H on the Far West Rand, there is very little doubt that the Boulder Reef of Luipaardsvlei Estates correlates

with the Big Pebble Conglomerate of the areas described thus far. It occurs only a short distance above the Kimberley Shales. On this basis the Battery Reef which is being mined on Luipaardsvlei Estates should correlate with the May Reef. The Battery Reef occurs only a few inches, or at the most a few feet, above the Boulder Reef.

In the Battery Reef the pebbles are up to 3 inches in diameter, while 12-inch boulders have been encountered in the Boulder Reef. The Battery Reef may thus represent a "re-worked portion" of the Boulder Reef.

(*b*) *West Rand Consolidated Mines*

The Boulder Reef of this mine is considered to be the equivalent of the Big Pebble Conglomerate. Cobbles in the Boulder Reef are up to 6 inches in diameter.

The economic horizon called the Pay Band or Battery Pay Band is considered to be the equivalent of the May Reef. It lies from 250 to 500 feet above the Kimberley Shales. It is a well-mineralized reef up to 15 feet in thickness, and contains large pebbles, often 3 inches in diameter. The highest values usually occur in the bottom band, which is associated with a carbon seam. It is, therefore, possible that the upper portion of the so-called Battery Pay Band actually corresponds to the May Reef Hangingwall Leader of the Eastern East Rand.

Near the common boundary with Luipaardsvlei Estates the Battery Pay Band occurs only about 2 feet above the Boulder Reef. In a westerly direction the two reefs diverge more and more until they are 60 to 70 feet apart in the south-western corner of West Rand Consolidated Mines.

The geological section of West Rand Consolidated South Shaft (Plate IX) illustrates the close correspondence between the West Rand and East Rand columns. In ascending sequence it appears that the following stratigraphic units can be recognised in this shaft: Bottom Reef, Chert-pebble Quartzite, Big Pebble Conglomerate, "Chloritoid Shale" Marker, May Reef and the "80" Foot Marker.

Pegg (1950) mentioned the presence of small chert pebbles in the quartzites above the Kimberley Shales in this area.

(*c*) *Randfontein Estates*

The Big Pebble Conglomerate is represented on this mine by the Lindum Reef. Actually the Lindum Reef is the bottom band of a wide conglomerate zone.

The Horsham Reef, considered to be the equivalent of the May Reef, lies at 110 to 125 feet above the Lindum Reef. It has an immediate footwall of either thin shale or soft grit, which presumably correspond to the "Chloritoid Shale" Marker of other areas. The stoped section of the Horsham Reef lies at the base of a 25-foot zone of conglomerates. The upper portion of the reef body is therefore considered to be the equivalent of the May Reef Hangingwall Leader of the Eastern East Rand.

The evidence therefore suggests the existence also in the West Rand area of the three major unconformities recognised within the Kimberley group of sediments in the East Rand Basin:

1. *Below the Bottom Reef.*

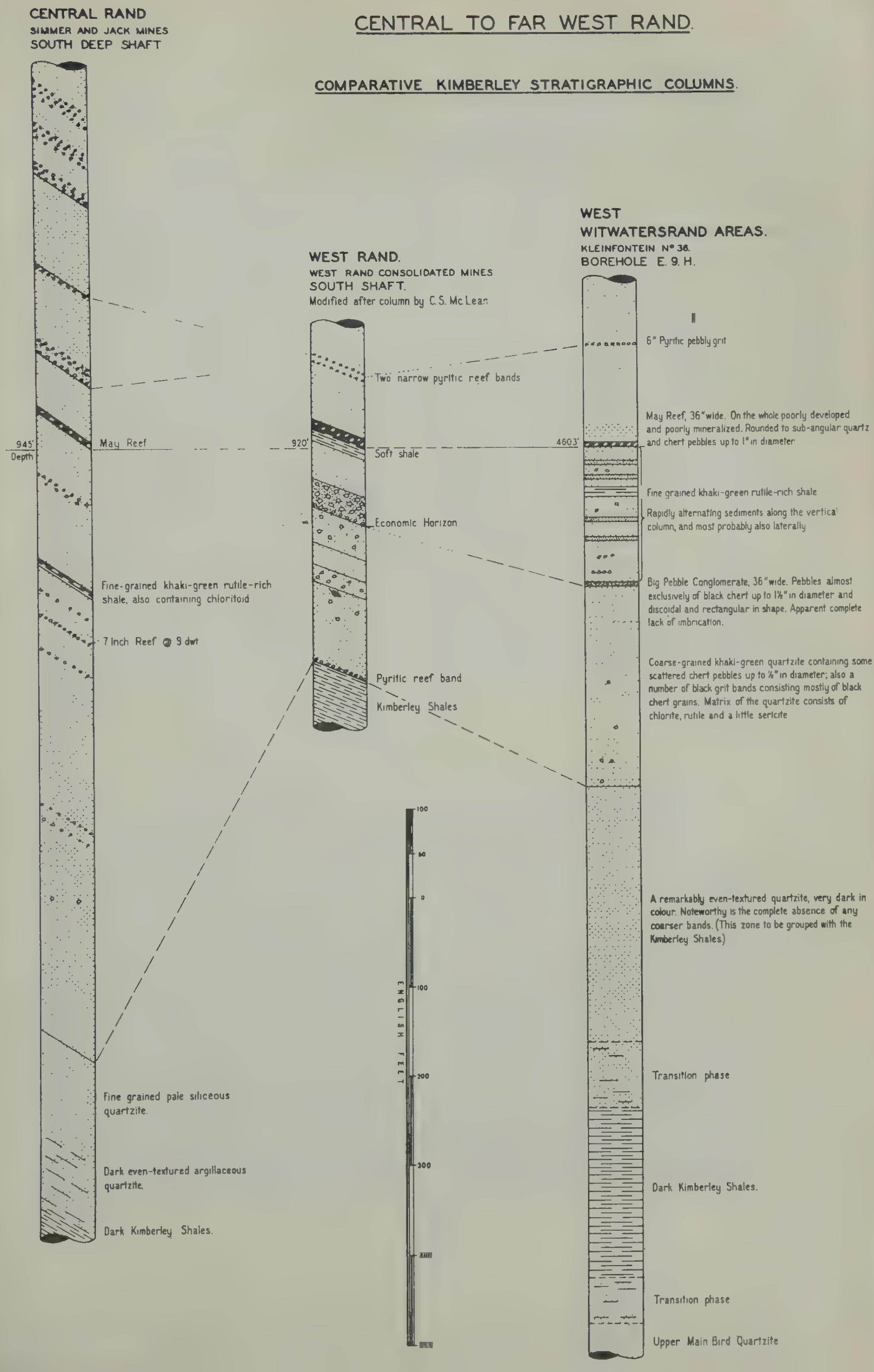
The upper quartzite phase of the Kimberley Shales is non-existent in the West Rand Consolidated South Shaft (Plate IX). Well to the east and to the south-west, in the Simmer and Jack South Deep Shaft and in borehole E.9H, this unit is again well developed.

2. *Below the Big Pebble Conglomerate.*

There is a very marked divergence between the Big Pebble Conglomerate and the Bottom Reef south-westwards from Luipaardsvlei Estates.

CENTRAL TO FAR WEST RAND.

COMPARATIVE KIMBERLEY STRATIGRAPHIC COLUMNS.



3. *Below the May Reef.*
The Big Pebble Conglomerate and the May Reef diverge in the same direction.

F. Klerksdorp Area

Borehole No. 2 on Witkop 46 and borehole No. 2 on Pilgrim's Estate 722 (Baines, 1949, Plate VIII) suggest the existence of the Kimberley Shales in the Klerksdorp area. A conglomerate which is locally known as the Gold Estates Reef most probably correlates with the Big Pebble Conglomerate of other areas, for three reasons:

1. It occurs in a suitable stratigraphic position to be the Big Pebble Conglomerate.

2. Beetz's description of the Gold Estates Reef as summarized by Baines (1949) fits the Big Pebble Conglomerate of other areas, except for the pebbles and boulders of "black mineralized shale". This summary reads:

> "a large pebble reef with intercalated layers of coarse quartzite; pebbles and boulders up to 8 inches in diameter of many different rocks, mainly quartz, chert, quartzite and black mineralized shale."

3. The Gold Estates Reef rests with strong unconformity on its footwall beds, according to Nel (1934), Baines (1949) and Beetz (1936). It appears that the erosion surface has cut right through the Kimberley Shales in some areas, also eliminating the Vaal Reef, which is the equivalent of the Basal Reef of the Orange Free State.

Baines (1949) states that the "upper portion" of the Big Pebble Conglomerate is being mined by New Klerksdorp Gold Estates Co., Ltd. The author suggests that these upper bands correlate with the May Reef. The relationship would then be the same as between the Battery Reef and Boulder Reef on Luipaardsvlei Estates.

We are apparently dealing, therefore, with at least two important unconformities in the Kimberley group of sediments in the Klerksdorp area, one below the May Reef and the other below the Big Pebble Conglomerate.

G. The Orange Free State Gold Field

De Kock (1951, p. 130) writes about the economic reefs in this field:

> "The principal economic reef horizons so far disclosed in the Free State are in order of importance: The 'Basal' Reef; the 'Leader' Reef; the 'B' Reef; the 'A' Reef; the 'Upper' Reefs and the 'Ventersdorp Contact' Reef. Of these the Basal Reef is by far the most important gold carrier in the Free State. Although it is too early for a final correlation the following tentative correlation of these reefs with those better known on the Rand is suggested. The Ventersdorp Contact Reef occurring at the base of the Ventersdorp Lower Lava is the counterpart of the same reef as known on Venterspost and Western Reefs. The Upper Reefs with their principal development in the Van den Heevers Rust area may be the equivalent of the Elsburg Reefs known in the Western Reefs mine. The 'A' Reef strongly resembles the 'May' Reef of the Kimberley Reef group of the East Rand. The 'B' Reef is closely similar to the lower Kimberley Reef which generally rests on the Kimberley Shales. The Leader Reef and Basal Reef can be correlated with reefs in the Bird Reef group, whilst there is now little doubt that the Basal Reef and the Vaal Reef occur on the same horizon."

Many of the conclusions arrived at in this chapter are the direct outcome of

frequent discussions with Dr. W. P. de Kock. His recognition of the May Reef in the Odendaalsrus area led the author to scrutinise the literature more closely. Dr. de Kock also arranged exchange visits between the author and Mr. G. W. S. Baumbach, resident geologist for New Consolidated Gold Fields in the Orange Free State. Mr. Baumbach recognised the close correspondence between the Vogelstruisbult Mine and certain parts of the Orange Free State gold field as regards the Kimberley group of sediments.

The areas to the south and south-east of Odendaalsrus are discussed separately.

(*a*) *The Area to the South-east of Odendaalsrus:*

A study of the literature on this area (Borchers and White, 1943) and Baines, (1949) and an examination of borehole cores (shown to the author by Mr. Baumbach) revealed that the group of sediments immediately above the Upper Shale Marker could be resolved into the fundamental units of the Kimberley group of sediments as known elsewhere. Correlation is thereby established. Reference to the average column by Borchers and White (1943, Fig. 1) is necessary to explain the details of the suggested correlation.

1. *Kimberley Shales*

The Upper Shale Marker is considered to be the equivalent of the Kimberley Shales. Borchers and White report that these beds are chloritoid-rich. A personal inspection of various boreholes did not confirm this.

2. *The Bottom Reef or "B" Reef*

A certain amount of erosion took place before the deposition of this reef, so that it may rest on either the Kimberley Shales or on the dull-grey quartzites which normally underlie the Kimberley Shales. It is of economic importance in some areas. When the Kimberley Shale Zone is absent owing to the stratigraphic break, the "B" reef can only be identified with certainty when the section above it is complete or nearly complete (Baines, 1949, Plate IX).

3. *The Chert-pebble Quartzite Zone*

This suite of sediments has been recognised in the area south-east of Odendaalsrus. It embraces an upper zone, "E.C.3 Elsburg Rare Conglomerates", and a lower zone "E.C.4. Elsburg Mixed Pebble Reefs".

The lower zone contains some robust conglomerate bands, which are probably highly lenticular. Otherwise the zone is a pebbly quartzite. Pebble types include white quartz, grey, black and banded chert, occasional red jaspers, and an unidentified khaki-yellow siliceous rock.

Generally speaking the upper zone is not very different from the lower. The quartzites of the upper zone are perhaps more argillaceous, while the pebbles may be somewhat smaller and more scattered. There appears, however, to be no need for a separation into an upper and a lower zone. This is one sedimentary phase. The beds are grouped together under the name suggested above.

4. *The Big Pebble Conglomerate* (*E.C.2 Elsburg Big Pebble Reef Marker*).

This horizon is a very important marker in the Orange Free State. The zone may be up to 180 feet thick. Pebble types include white quartz (quite often predominating), black and banded chert, while occasional shale pebbles have been encountered. The pebbles are usually up to 2 or 3 inches in diameter, except towards the base, where they may reach a maximum diameter of 6 inches.

The Big Pebble Conglomerate is not always readily identified in borehole cores, for three reasons:

(i) It may change laterally into a quartzite zone, thereby making recognition difficult.

(ii) In some areas the zone is entirely eliminated by the unconformity below the May Reef.

(iii) Similar robust conglomerates may appear in the suite designated the "Chloritoid Shale" Marker. The upper limit of the Big Pebble Conglomerate can therefore not always be determined with certainty.

There is positive evidence that the Big Pebble Conglomerate covers an erosion surface (Baines, 1949, p. 313). This is additional evidence in favour of the correlation put forward in this account.

5. *The Puddingstone Reef*

This conglomerate was recognised in one borehole from the area south-east of Odendaalsrus. It occurs in the correct stratigraphic position. Isolated pebbles of about ¾ inch in diameter occur, embedded in a shaly matrix. The core of this borehole was shown to the author by G. W. S. Baumbach.

This peculiar conglomerate is apparently intermittently developed in the sector under review. It has not been described in any of the publications.

6. *The "Chloritoid Shale" Marker*

Typical chloritoid-rich shale sometimes occurs above the Big Pebble Conglomerate. The tiny chloritoid crystals can be observed in hand-specimen, scintillating with numerous pin-points of light. It is noteworthy that the Kimberley Shales (Upper Shale Marker) generally do not display this characteristic. The reason for this fundamental difference between the two shales is not clearly understood.

In areas where the typical chloritoid-rich shales are not developed, the "Chloritoid Shale" Marker may be recognised in a different form, as is also the case in the East Rand Basin. The "E.C. 1 Elsburg Conglomerate Zone" of Borchers and White appears to be the equivalent of the "Chloritoid Shale" Marker. It is largely composed of an argillaceous quartzite with a distinct yellowish tinge. This characteristic colour is almost certainly due to an abundance of the small rutile crystals. The zone includes a number of conglomerate bands.

7. *The May Reef.*

The "A" Reef (Baines, 1949, p. 311) which occurs on the average about 70 feet above the Big Pebble Conglomerate, is considered to be the equivalent of the May Reef.

It is not always readily identified for the following reasons:

(i) In some areas it occurs sandwiched between other conglomerates.

(ii) It is of lense-like character according to Baines. It may thus apparently degenerate into a mere pencil-line contact. An intimate knowledge of the hangingwall and footwall rocks would then be necessary in order to recognise the horizon.

(iii) A high gold content would assist to single out the horizon.

Features of the reef are the dark matrix and the small smoky quartz pebbles. Occasionally it rests on chloritoid-rich shales, when correlation offers no difficulty.

8. *The* "80" *Foot Marker*

The "V.S. 5 Ventersdorp Basal Agglomerate-Conglomerate" of Borchers and White (1943) appears to be the equivalent of the "80" Foot Marker of other areas.

The V.S. 5 may develop into a zone of conglomerates up to 100 feet thick. At the other extreme, it is often represented by a single narrow pebble bed.

9. *The Elsburg Group of Conglomerates*

It is possible that the "V.S.1 Ventersdorp Agglomerate-Conglomerate" corresponds to the Elsburg conglomerate group of the Rand. The zone has an average thickness of between 700 and 800 feet.

(*b*) *The Area to the South of Odendaalsrus—the St. Helena Sector*

The information for this area has largely been obtained from a publication by Frost and others (1946).

Reference to Plate VII of this publication is necessary.

After consideration of the detailed section of the "St. Helena Reef Zone", the following remarks are offered on the subject of correlation:

1. The correlation of the "St. Helena Reef Zone" as a whole with the "Upper Elsburg Reef Zone" can now be revised.

2. The sediments from the base of the so-called "Ventersdorp Basal Conglomerate" down to and including Frosts' "Leader Reef" are tentatively correlated with the Kimberley group of conglomerates, the evidence being:

(i) The Big Pebble Conglomerate is now known and accepted as a regional marker, and undoubtedly belongs to the Kimberley conglomerate group.

(ii) The narrow persistent reef shown about 12 feet above the Big Pebble Conglomerate occurs in a stratigraphic position that is suitable for the May Reef.

(iii) Lithologically the Mixed Pebble Conglomerate Zone compares well with the Chert-pebble Quartzite Zone of other areas (E.C. 3 and E.C. 4 of Borchers and White.)

(iv) The "B" Reef occurs at the base of the Chert-pebble Quartzite Zone. That would indicate that Frost's "Leader Reef" is not the same horizon as the Borchers-White "Leader Reef". The former appears to be the Bottom Reef of the Kimberley group, while the latter is considered by de Kock (1951) to be a member of the Bird Reef group. The explanation is found in the stratigraphic break immediately below the Bottom Reef, which resulted in the elimination of the Kimberley Shales and the uppermost Bird Reefs. The Bird Reefs therefore apparently contributed towards the mineral content of the "B" Reef in some areas.

3. It is possible that the Kimberley Shales have been preserved in some small areas in the sector under review.

4. It therefore follows that the Basal Reef and the Leader Reef proper should fall within the Main-Bird Series, apparently Upper Main-Bird, and may therefore be members of the Bird Reef group.

5. The correlation of the so-called "Ventersdorp Basal Conglomerate" with the Elsburg conglomerate group is a possibility.

(*c*) In support of the above analysis, the following passages are quoted from Baines and Borchers.

1. Baines (1949), page 310:

"It should be remembered, however, that the Leader Reef of the St. Helena Area is called the 'B' Reef elsewhere, and that the Leader Reef of elsewhere does not occur in the St. Helena Area borehole cores except possibly in some instances in close proximity to the 'B' Reef."

2. Borchers (1946), page 29:

"Again, St. Helena is unique in that no shales are developed in the Reef Zone whereas, almost without exception, everywhere else either the Khaki shale (above the Basal Reef) or the Upper Shale Marker or both, and at times a third shale horizon, are strongly developed and of very considerable importance as Marker Horizons."

H. The Vredefort Area

Sharpe (1949) reported chloritoid-rich sediments from the Kimberley beds in the Great Western Mine near Vredefort.

XIV THE DEPOSITIONAL ASPECT

A. Continuity of Horizons

It has been shown that the Kimberley-Elsburg Series or portions of it can be recognised over the bigger part of the large structural basin defined in the introductory pages. Furthermore, it has been shown that a regional correlation of individual stratigraphic units of the Kimberley-Elsburg Series can be effected.

The sedimentary column and the lithological characteristics of these rocks in the East Rand Basin have been used in recognising the Series in other areas. Except for minor differences of which there may be many, the broad pattern indicated is considered to be substantially correct. One has to review the entire structural basin in order to arrive at a better perspective of some aspects of the problem.

The remarkable continuity of certain horizons over large areas, and the constancy of the lithological character of these horizons, concerning both coarse and fine sediments, assist in long-distance correlation, and have contributed to the rapid expansion of the Orange Free State gold fields. Of importance in this respect are:

(*a*) The Kimberley Shales, which are regarded by the author as the upper phase of the Main-Bird Series.
(*b*) The Chert-pebble Quartzite Zone.
(*c*) The Big Pebble Conglomerate.
(*d*) The "80" Foot Marker, and
(*e*) zones of clear quartzites.

In an endeavour to determine the environment(s) and modes of deposition of the sediments of the Kimberley-Elsburg Series, certain environments can thus be eliminated at the outset.

The remarkable persistence regionally of such a coarse unit as the Big Pebble Conglomerate indicates very powerful transporting and distributing agents. Furthermore, its very constanct lithological character indicates:

1. The same set of conditions controlling its deposition over a very large area, and/or
2. derivation of the constituent materials from a lithologically constant terrane.

The units of clear quartzites also have very wide distribution, indicating:

1. Constant type of material supplied, and/or
2. constant conditions of deposition over large areas.

There is one important exception to the above general rules: the suite of sediments designated the "Chloritoid Shale" Marker, which exhibits a very marked change of facies. In the East Rand Basin the change from coarse to fine takes place from west to east, or probably more correctly, from north-west to south-east.

The lateral change of conglomerate zones into clear quartzites (e.g. the upper Kimberley conglomerate clusters in the East Rand Basin) conforms to a perfectly normal behaviour of coarse clastics the world over. Such a feature indicates the direction from which the sediments came, and is not in the first place a pointer to a specific environment of deposition.

Ilmenite was continuously available during the time of formation of the Kimberley Shales, as indeed it was during the period of formation of the entire Upper Witwatersrand System, as indicated by the alteration products leucoxene and rutile. All three minerals occur in the Jeppestown Shales immediately below the Main Reef in the East Rand Basin.

Of paramount importance is the fact that certain stratigraphic breaks instituted within the succession assume a regional character. Economically it is very important to recognise these regional surfaces of erosion. Not only do they assist in long-distance correlation, but they also have a very important bearing on the economic possibilities of certain reefs. Only detailed stratigraphic study made their recognition possible. The May Reef could be related to older (parent) gravels. The same remark apparently applies to the "B" Reef in certain parts of the Orange Free State gold field, notably in the area immediately to the south of Odendaalsrus. A knowledge of the sum total of the gold contained in the source gravels is of prime importance in assessing the economic possibilities of the May Reef and "B" Reef in those areas where the erosion surfaces cut into the older gravels. The May Reef represents the heavy, resistant and coarse fraction of a large volume of sediments.

The sub-May-Reef erosion surface always has a low relief, but the erosion surface on which the Big Pebble Conglomerate rests has considerable relief locally. The unconformity below the "B" Reef is apparently also one of low relief.

B. A Discussion of the Environments Considered

This discussion is based on the classification of environments by Twenhofel (1932 and 1939).

It is conceivable that the sediments of the Kimberley-Elsburg Series could have formed partly under one and partly under another environment. The possibility that the sediments of the Kimberley-Elsburg Series occupied dry land at one or more stages has been investigated.

The following environments have been given consideration:

(*a*) *Continental Environments*

Of these the desert environment requires no comment, as there is no evidence to support the deposition of even a portion of the Kimberley-Elsburg Series in this environment.

Considering the abundance of gravel that was necessary for the formation of the Big Pebble Conglomerate, the possibility of *glacial* accumulations in the piedmont and lower piedmont cannot be ignored. It is, however, not suggested that the Big

Pebble Conglomerate was deposited by glacial action on a piedmont plain. The gravels might have been produced in part and brought nearer to the site of permanent deposition by glacial action.

The aqueous environments piedmont, valley flat and lacustrine are discussed.

(*b*) *Transitional Environments*

The littoral and delta environments have been given consideration.

(*c*) *Epicontinental Seas and the Marine Neritic Environment*

There is positive proof that portions of the Kimberley-Elsburg Series have been deposited in the marine neritic environment.

(*a*) *Continental Environments*

1. *The Piedmont Environment*

Important characteristics of these deposits are:

(i) The types of deposits range from alluvial cones and fans to alluvial screes on steep slopes.

(ii) Original inclinations of units are therefore expected to be high.

(iii) Sorting and stratification are usually imperfect, also in the lower reaches of the environment, where it grades into that of the valley flat. Extensive continuity of individual horizons is therefore not to be expected, nor regional surfaces of erosion of low relief. A piedmont environment for the sediments under review is therefore eliminated.

The great Siwalic System of Northern India was partly deposited in this environment.

2. *The Valley Flat Environment*

Because of lithological characteristics the following suites of sediments are not incompatible with the flood plain environment.

(i) The "Chloritoid Shale" Marker.

(ii) The Chert-pebble Quartzite Zone.

(iii) Zones of clear quartzites.

But these sediments can also be reconciled with the marine neritic environment.

The deep erosion in portions of the Eastern East Rand which resulted in the elimination of the entire Kimberley Shale Zone along limited zones, must be ascribed to processes of the valley flat environment.

Sea currents can act to surprising depths and may seriously interfere with sedimentation processes. It is not clear to what extent they can erode a bottom. Along narrow passages between islands one could expect destructive work to be accomplished and the creation of hard bottom at considerable depths. But in open neritic waters it is extremely doubtful whether a marine current could create scour depressions comparable to those on Marievale and in the May Shaft area on East Daggafontein.

In the flood plain environment the streams have lesser capacity and competency as compared to the piedmont environment. A regional horizontal persistence of individual horizons, especially if composed of coarse clastics, is therefore not to be expected. Extensive sheets of gravel, although possible, are generally not repeated many times in the same system of flood plain deposits. The author considers that the Big Pebble Conglomerate cannot be reconciled with flood plain deposits. At the same time it has been shown that the Big Pebble Conglomerate was the initial deposit in the scour depressions of the Eastern East Rand, occurring also outside these

depressions over large areas on an apparently even floor. This relationship will be discussed in more detail in the final chapter.

The Siwalic System of Northern India was partly deposited in this environment.

Reinecke (1930) parallelled the Witwatersrand System with the Siwalic formation in the general lithology of the stratigraphic column and environments of deposition. His arguments are very convincing, except for the one fact that nothing like the extensive gravel sheets of the Witwatersrand System (Government Reef, Main Reef, Main Reef Leader, Big Pebble Conglomerate, May Reef, etc.) have been described from the Siwalic.

3. *The Lacustrine Environment*

The evidence produced here suggest that the sediments of the Kimberley-Elsburg Series were deposited, at times, in an environment characterized by strong waves and currents.

Positive and negative movements of shore lines must have been important in controlling the production and deposition of some of these sediments.

As strong waves and currents generally do not develop in relatively small, relatively shallow enclosed bodies of water, the author also eliminates the lacustrine environment.

(*b*) *Transitional Environments*

1. *The Littoral*

This is the part of the shore zone exposed at low tide and flooded at high tide. The sediments of the littoral environment (boulders to clays) seldom attain permanent deposition. They are of little quantitative importance in the preserved sediments of the world.

2. *The Delta Environment*

The component parts of deltas have not been recognised. There is no evidence to support the deposition of even a portion of the Kimberley-Elsburg Series in this environment.

(*c*) *The Marine Neritic Environment:*

The May Reef could have formed in one of two ways:

1. As a basal conglomerate developed during a period of positive advance of a wave front relative to the land, or
2. Concentrated from material eroded from a neritic floor, when the latter had been built to a base level of deposition, and subsequently raised and in parts warped to above that level.

The sediments for some distance above the May Reef can thus be classed as neritic.

XV CONCLUSION

During the past ten years a great deal of information has accumulated concerning the sediments of the Upper Witwatersrand System. The discoveries in the Orange Free State and Klerksdorp areas have stimulated interest, and it is today generaly realized that detailed stratigraphic study is necessary for correlation purposes. Long-distance correlation is possible, and has been facilitated by the remarkable continuity of individual horizons. A fairly comprehensive picture exists of the distri-

bution of individual reefs and suites of sediments, and of their lithological character and inter-relationships. That much has been achieved. The complete reconstruction of the environment(s) and mode of deposition of the sediments of the Upper Witwatersrand System is not yet possible. In the case of the Kimberley-Elsburg Series certain positive deductions can be made as indicated in the previous chapter. These conclusions will be discussed in more detail in this chapter.

It has been shown that three regional unconformities exist within the section extending from below the Kimberley Shales to above the May Reef (Plates I and II).

The Kimberley-Elsburg Series will be discussed in four stages based on these three regional unconformities:

A. The Section from the May Reef to the Top of the System

This includes the upper Kimberley conglomerate clusters, the Kimberley-Elsburg Intermediate Quartzites and the Elsburg conglomerate group (Plates I and II).

B. The Middle Kimberley

This includes the section from the base of the Big Pebble Conglomerate Zone to the top of the Micaceous-gritty Quartzite (Plates I and II, and Table I).

C. The May Reef

D. The Lower Kimberley

This includes the "B" Reef and the Chert-pebble Quartzite Zone.

A. The Section above the May Reef

It has been shown in the previous chapter that for some distance above the May Reef the sediments can be assigned to the marine neritic environment. As no regional breaks in the sedimentation have been discovered in the section from the May Reef to the top of the System, there is no reason to assign any portion of this section to another environment.

The thick conglomerate clusters in the section under review have been used by other investigators (notably Reinecke, 1930) as evidence against a marine environment. There is, however, one important feature which argues in favour of the marine environment. The "80" Foot Marker in the East Rand Basin provides this evidence. This conglomerate cluster is in the form of a zone of pebbly quartzites and lenticular conglomerates, capped by a well-developed fairly persistent coarse conglomerate (Plates I, II, V 3 (*a*) and Fig. 5). This upper band is frequently the coarsest band in the cluster under review. In the author's opinion this feature would indicate a building up of the depositional floor to a profile of equilibrium. The uppermost coarse continuous band would then represent a layer concentrated from a considerable volume of sediments at the level of the profile of equilibrium as the associated finer sediments were carried into deeper waters. In this respect it is interesting to remember that the top band of the "80" Foot Marker gave sporadic payable gold values on Government Gold Mining Areas.

The higher clusters (Sharpe's U.K.5 and U.K.3) were not examined in detail for similar features in the East Rand Basin. But a certain band in the U.K.3 cluster

gave payable gold values on Government Gold Mining Areas. It is possible that it owes its higher gold content to the same sequence of events as described above.

Recently the author had an opportunity of studying the core from four boreholes which penetrated these upper clusters on the Simmer and Jack Mine on the Central Rand. It was possible to recognise five such coarse persistent bands, in each case capping pebbly quartzites and apparent lenticular conglomerates in which the average pebble size was much less.

It is possible that similar features exist in the Elsburg conglomerate group.

B. The Middle Kimberley

(*a*) *The Big Pebble Conglomerate*

This conglomerate zone attains a maximum thickness of 160 feet in the East Rand Basin and 180 feet in the Orange Free State gold field. In the Klerksdorp area thicknesses up to 140 feet are known.

The Big Pebble Conglomerate covers an erosion surface which has considerable relief locally, e.g. in the May Shaft area on East Daggafontein and in the central portion of Marievale (Fig. 3). Whiteside (1950) regarded the deep erosion in the May Shaft area as post-Big Pebble Conglomerate. But on Marievale the conglomerate zone that covers the base and sides of the scour depression is so similar to the Big Pebble Conglomerate, and the overlying argillaceous zone so similar to the "Chloritoid Shale" Marker of surrounding areas, that the author has no choice but to date the deep erosion in this area as immediately pre-Big Pebble Conglomerate. Outside these depressions the Big Pebble Conglomerate covers a fairly even (eroded) surface, as far as can be judged from boreholes and the underground workings. According to Beetz (1936) the Gold Estates Reef (i.e. the Big Pebble Conglomerate) appears to have been deposited in scour depressions in portions of the Klerksdorp gold field.

The Big Pebble Conglomerate is incompatible with a flood plain environment, while the scour depressions which it partly fills in some areas can be ascribed only to sub-aerial erosion.

In the author's opinion the conditions that led to the formation of the Big Pebble Conglomerate are therefore as follows:

1. An initial land surface built largely of the Chert-pebble Quartzite Zone.
2. Erosion of this land surface by rivers to considerable depths along narrow belts resulting in the elimination of the entire Chert-pebble Quartzite Zone, the entire Kimberley Shale Zone and even portions of the Bird Amygdaloid (Fig. 3).
3. Accumulations of enormous volumes of gravel in the higher reaches of this land-eroded surface.
4. Rapid submergence (drowning) of a large tract of country, resulting in the preservation of the land topography beneath the sea.
5. The sea was thus brought into contact with the accumulations of gravel as suggested above; while strong waves and currents which now prevailed in the invaded territory were responsible for the distribution of the coarse material to build the Big Pebble Conglomerate.

This sequence of events would explain the author's view of the Big Pebble Conglomerate forming the initial deposit in the deep scour depressions, partly filling them, and at the same time extending for great distances outside these depressions on a more even floor. We are thus apparently dealing with a case of drowned topography,

smothered by marine sediments. The depressions were further filled up by muds and sands (the "Chloritoid Shale" Marker).

(*b*) *The "Chloritoid Shale" Marker*

The basal members of this suite were deposited on a fairly even floor, as the Big Pebble Conglomerate is seldom absent from below this suite (Plates III, IV, V 1 and VI). The Puddingstone Reef is regarded as a transition phase between the Big Pebble Conglomerate below and the rutile-rich shales above.

This suite can thus be regarded as following conformably above the Big Pebble Conglomerate, and may thus also be regarded as having been deposited in the sea. The change from coarse sands and gravels to fine-grained sands and muds in the East Rand Basin conforms to the general behaviour of marine gravels whereby the coarser constituents were deposited slightly earlier than the associated finer materials.

As previously described, there is evidence of a certain amount of scouring and re-deposition during this period. Lanes of scour could have been produced by off-shore currents, resulting in streak-like intervening deposits. In the Sub Nigel No. 3 Shaft area a lense of arenaceous shale apparently fills a shallow scour depression in conglomeratic quartzites.

The irregular fragments of shale embedded in quartzite and quartzitic shale near the base of the suite might suggest local piedmont conditions or undermining of river bands on a flood plain. As far as can be judged from the underground workings, however, the beds containing these shale fragments have been deposited on a very even floor. It would thus appear that they correspond to Twenhofel's (1932) "intra-formational conglomerates". On page 216 he writes:

> "As defined by Walcott, these are conglomerates developed by the breaking up of a partially consolidated bed and the incorporation of the fragments in new strata nearly contemporaneous with the original beds. Such are known to be developed under marine conditions, where partially consolidated materials are torn up by strong waves and redeposited. Laminae and bedding planes arch downward between the fragments and these lie in all sorts of positions."

(*c*) The "Upper Parent Conglomerate" of the May Reef occurs immediately above the "Chloritoid Shale" Marker in the Western East Rand (Plate I). The author suspects that it also exists on the Simmer and Jack Mine on the Central Rand. It is not known to what extent it existed in the Eastern East Rand, but judging by the map of the May Reef floor (Plate VII), it appears that it must have existed in at least a portion of the eastern sector. It has not been encountred in any other area.

A certain amount of erosion took place before its deposition in the northern portion of Brakpan Mines. Elsewhere in the East Rand Basin there is no positive proof of an unconformity. For practical purposes it is thus regarded as following conformably above the "Chloritoid Shale" Marker. It is difficult to reconcile it with the flood plain environment.

C. The May Reef

It has been thought best to discuss the May Reef at this stage.

As has been indicated above, there is strong evidence for regarding the May Reef footwall sediments down to and including the Big Pebble Conglomerate as marine deposits. According to Whiteside's interpretation of the scour depression in the May Shaft area, the origin of the May Reef can be ascribed to the invasion of a land area

by the sea. But the author has shown that these scour depressions should be dated as immediately pre-Big Pebble Conglomerate. The author believes that the May Reef was formed by the erosion of a neritic bottom, when the latter was raised regionally and arched upward over fairly large areas. The bottom was thus brought into a position where it could become subject to erosion. Where the floor was of such a nature as to produce gravel, the coarse and heavy fraction of the eroded material was concentrated on that part of the bottom that was eroded, and finally deposited in the immediate vicinity on the sub-May Reef synclines (Plates V 1, VI and VIII). Stratigraphers have favoured this type of unconformity very little in the past, as so very little is known of the erosion processes at and above the base level of deposition. In this connection Twenhofel (1939) writes:

> "The fact of wave activity affecting neritic bottoms indicates that these are nearing, and perhaps have reached, the level of the base level of deposition, so that in many instances the sediments have but temporary deposition."

To contrast the above concept with the landward migration of shore lines, the following passages are quoted from Twenhofel (1932), where he discusses the sediments produced by—

1. an invading shore-line with stationary sea level, and
2. an advancing sea caused by a rise of sea level.

This excludes extremely rapid migration resulting from the submergence of large tracts of country. These quotations are:

1. "An invading shore-line with stationary sea level develops an eroded bottom surface coextensive with the area of invasion. No permanent deposits are made on this surface so long as sea level remains unchanged. Materials torn from the rocks of the coast remain with the beach or in shallow water until reduced to particles sufficiently small to be transported across the wave-cut surface to deeper waters below the profile of equilibrium. The sequence should contain few or no coarse deposits."
2. " . . . basal conglomerates are possible under the conditions of an advancing sea due to rise of sea level other those portions where the profile of equilibrium is above the new bottom, and are not probable under other conditions. As basal conglomerates appear to be rather rare at the bases of marine sequences, it therefore appears that the inland migration of most shores must have been very slow instead of rapid."

Considering the intimate relationship between the May Reef and the eroded anticlines and synclines (Plates V, VI, VIII and Fig. 6), and considering at the same time the above conclusions by Twenhofel, the author finds it impossible to ascribe the origin of the May Reef to the landward migration of a shore line, under any of the conditions outlined above.

Another important consideration is the fact that the gold and associated heavy minerals are generally concentrated at and near the summit of the May Reef (Plate V, 4 (*c*)). In the case of a basal conglomerate resulting from an advancing wave front, the heavy minerals are most likely to be concentrated at and near the base of the conglomerate. As too little is known at present of the erosion processes at and above the base level of deposition in the marine neritic environment, the origin of the feature under review remains, for the present, unexplained. It is, however, conceivable that, when a neritic floor is in a position at or above a profile of equilibrium or the permanent base level of deposition, the sediments eroded from the neritic floor would be shifted back and forth, the finer and lighter fraction ultimately escaping to deeper waters. In the case of the May Reef it could have been possible, by such action, for

the fine-grained heavy minerals to have been separated from the pebbles, and finally washed on top of the latter (Plate V, 4).

D. The Lower Kimberley

This part of the column embraces the "B" Reef and the Chert-pebble Quartzite Zone.

The "B" Reef covers an erosion surface. In the East Rand Basin this reef is intermittently developed, and lithologically does not differ from the lenticular conglomerates associated with the overlying Chert-pebble Quartzite Zone. In the East Rand Basin the constituent materials of the "B" Reef were not obtained from the floor on which it rests, but in the Orange Free State gold field the erosion surface cuts through the Kimberley Shales in some areas, eliminating also some of the upper Bird Reefs. These reefs therefore apparently contributed to the mineral content of the "B" Reef in the Orange Free State.

It has not been established whether the unconformity is due to erosion of a neritic floor, or due to erosion by an invading sea. Because of these two possibilities, the Chert-pebble Quartzite Zone could therefore also be classed as a marine neritic deposit. But it has been shown that these sediments occupied dry land before the deposition of the Big Pebble Conglomerate. We are therefore apparently dealing with a withdrawal of the sea after the formation of the Chert-pebble Quartzite Zone.

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PLATE X

Ph. 1. May Reef on chloritoid shale. Note predominance of white vein quartz pebbles. Scale: 9″ rule.

Ph. 2. 3. May Reef: an intermittent line of large chert pebbles.
2. Puddingstone Reef.
1. Chert-pebble quartzite.
Note: (i) angular difference between May Reef and its footwall beds, and (ii) big Pebble Conglomerate non-existent. Scale: 9″ rule.

Ph. 3. 4. May Reef Hangingwall Leader.
3. Siliceous Quartzite Bar.
2. May Reef Horizon: almost a barren surface.
1. Puddingstone Reef. Scale: 9″ rule.

Ph. 4. 10-inch robust May Reef in direct contact with the Puddingstone Reef. Scale: 9″ rule.

Ph. 5. 10-inch dark May Reef resting on a grey arenaceous shale. Pebbles of white and smoky quartz up to $\frac{3}{4}''$ in diameter. Scale: 9″ rule.

Ph. 6. 2. May Reef.
1. Big Pebble Conglomerate. Bottom of rule at junction. Note the black chert pebbles in the lower horizon. Scale: 9″ rule.

Ph. 7. Very wide May Reef in a stope—chloritoid shale footwall. Note predominance of small white vein quartz pebbles; very few chert pebbles present.

Ph. 8. 5. Siliceous Quartzite Bar. Note isolated pebble.
4. May Reef simulating the Big Pebble Conglomerate in appearance. Note large chert pebbles.
3. Quartz ("Chloritoid Shale" Marker).
2. Puddingstone Reef.
1. Clusters of pebbles in the Chert-pebble Quartzite. Big Pebble Conglomerate non-existent. Scale: 9″ rule.

PLATE X

PLATE XI

Ph. 9. May Reef horizon has transgressed on to the Puddingstone Reef: almost a barren surface. Scale: 9″ rule.

Ph. 10.
4. Dark hangingwall quartzite.
3. May Reef horizon: an intermittent line of pebbles.
2. Big Pebble Conglomerate, partly represented by quartzite.
1. Chert-pebble Quartzite.
Note transgression by the May Reef. Scale: 9″ rule.

Ph. 11. May Reef horizon has transgressed on to the Chert-pebble Quartzite. No deposition of May Reef took place here. Note the dark hangingwall quartzite. Scale: 9″ rule.

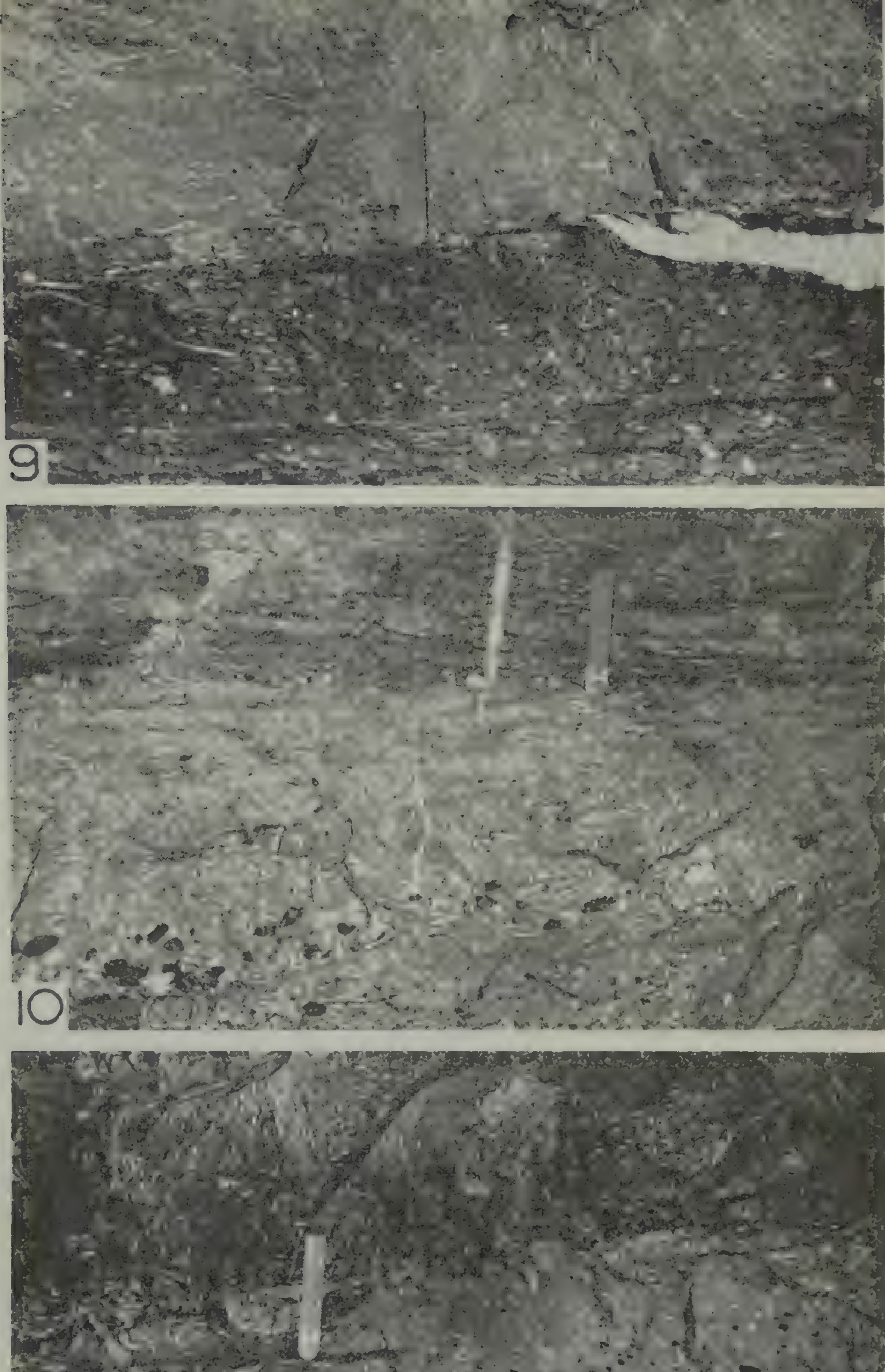

PLATE XI

THE RIVER EVOLUTION AND THE REMNANTS OF THE TERTIARY SURFACES IN THE WESTERN LITTLE KAROO

By

C. J. LENZ, B.Sc.

(Submitted March, 1953)

ABSTRACT

In this essay an attempt has been made to establish the relationship between surface evolution of landforms, regional and local geological structure, and the localisation of stream channels. The various remnants of stream sculpture, surfaces were mapped and correlated, and the nature of the deposits covering them were studied with the view to establish the climatic conditions which obtained during various epochs of the Tertiary Period. It was also attempted to establish the effects of epeirogenesis on stream rejuvenation in the area concerned and to present a reconstruction of the development of the Gouritz River System.

This paper was awarded the third prize of the Geological Society of South Africa for the year 1954

CONTENTS

I LOCATION AND TOPOGRAPHY

The area investigated includes the magisterial districts of Touws River, Montagu, Swellendam, Laingsburg, Ladismith, Riversdale and Calitzdorp. It is situated in the Little Karroo and the Southern Folded Belt. The dominant feature of this area is the sub-parallel mountain ranges and the intermontane lowland belts which run approximately east-west, being gently arcuate to the south. In this area the northern range consists of the Witteberg and Elandsberg, backed, a few miles to the south, by the Swartberg Range. This range consists of the Aynsberg in the west, and the Klein Swartberg and the Great Swartberg in the east. This is by far the highest range in the area, rising to over 7,500 feet above sea level near Ladismith.

Between this range and the southernmost one—the Langeberg—there are a number of smaller, more or less isolated mountains. These are Touwsberg, Waboomsberg, Rooiberg (and the Gamka Hill) and the Warmwatersberg, rising from 2,000 to 4,000 feet above the surrounding country.

The Langeberg Range to the east of Gouritz Poort is locally known as the Attaquas Mountains. In the west this range, which trends east-west in the south, swings to the north-west and is joined to the Swartberg Range by the Waboomsberg and the north-east—south-west trending Couga and Nouga Mountains. This feature, together with the fact that the Rooiberg-Gamka Hills also make a slight angle with the southern range, gives the western portion of the area a canoe-shaped appearance. This is called the Ladismith Karroo, and the adjoining area to the east, bounded in the east by the Kamanassie Mountains, is termed the Oudtshoorn Basin.

II PROBLEM, METHODS AND ACKNOWLEDGEMENTS

In his book "A Glimpse of South Africa", Prof. M. S. Taljaard (1949) admits that the views he expresses regarding the drainage evolution in this area, are based on visual observations alone, and that more certainty can only be expected after an "exhaustive investigation and mapping programme has been completed". The writer has attempted this in the portion of the area outlined. He would here like to express his gratitude to Prof. Taljaard for his guidance and interest during the work. To Prof. D. L. Scholtz the writer is also indebted for help and advice. Of the many farmers in the area to whom thanks are due, the writer wishes to mention the late "Oom Jan" (J. J. M.) van Zyl and Mrs. van Zyl in particular for their kindness.

The solution to the problem of the drainage evolution clearly lies in determining the ages of the breachings of the mountains by the rivers. As a basis certain of the Tertiary surfaces were mapped and certain heights determined. The distribution and forms of the remnants of these surfaces were taken from aerial photographs, and in this connection the writer would like to thank the Director of the Trigonometrical Survey (Mowbray) for the use of a number of their photographs, and Mr. P. W. Thomas of that office, in particular, for his help.

The best topographic map of the area that could be obtained had a scale of 1 : 250,000 which proved to be inadequate. During 1949 and 1950 a considerable amount of work was done, using telescopic alidade and Paulin altimeter. However,

the lack of tertiary triangulation beacons in a large part of the area made this method too slow and inaccurate. In 1952 a certain amount of the area was resurveyed with the aid of a barograph at base camp to check the altimeter. Readings were, wherever possible, confined to the mornings and within a radius of 10 miles of the base camp on the same side of the mountain, and checked against known beacons. By this method it is hoped that the possible error was reduced to less than 25 feet, although readings were taken to 5 feet. As Johnson (1944) has pointed out, the only reliable readings are those where the surface abuts against the mountains. Slumping and the fact that the remnants are nearly always discontinuous with the mountain, made accurate determination difficult. However, there are few surfaces in the area and thus correlation is, for the most part, relatively simple as the difference in elevation is usually a few hundred feet.

Because of the size of the area (50×115 sq. miles) it was impossible to study all the remnants of the surfaces thoroughly.

III STRATIGRAPHY, STRUCTURE AND HISTORY

The area displays part of the succession, conformable in the south, which stretches from the Palaeozoic into the Mesozoic era and has been divided into the trinal Cape System and the Karroo System (Fig. 1). These formations have been folded, intensely in the south and dying out towards the north. In the eroded syncline of the Oudtshoorn Basin, and to the south of the Langeberg, Cretaceous (Enon) rocks rest on steeply dipping Bokkeveld and Table Mountain Series, with a low dip to the north. To the north they are bounded by normal faults which have been instrumental in bringing pre-Cambrian sedimentary and granitic rocks to the surface. Melilite basalt has intruded the Enon deposits in the south. Resting almost horizontal over all these rocks are remnants of surfaces bearing deposits of Tertiary and younger epochs.

The sequence of the main events in the history of the Fold Belt is given below.

Planation of pre-Cape formations and deposition of Cape and lower Karroo Systems.	Ordovician. Devonian. Carboniferous.
Folding, in pulses (main pulse pre-Molteno).	Permian.
Continued deposition in north (Stormberg. Series).	
Erosion in south.	
Outpouring of Drakensberg Lavas.	Triassic
Erosion.	Jurassic.
Sagging, Crossfolding, Faulting and Deposition of the Enon Deposits.	Mid- } Late- } Cretaceous.
Differential uplift and melilite basalt injections.	
Formation of higher surface and deposits.	Tertiary (Eocene-Oligocene)
Warping (tilting and uplift).	Tertiary (Mio-Pliocene).
Formation of lower surface gravels.	
Further uplift.	
Recent fluctuation of sea level.	Quaternary.

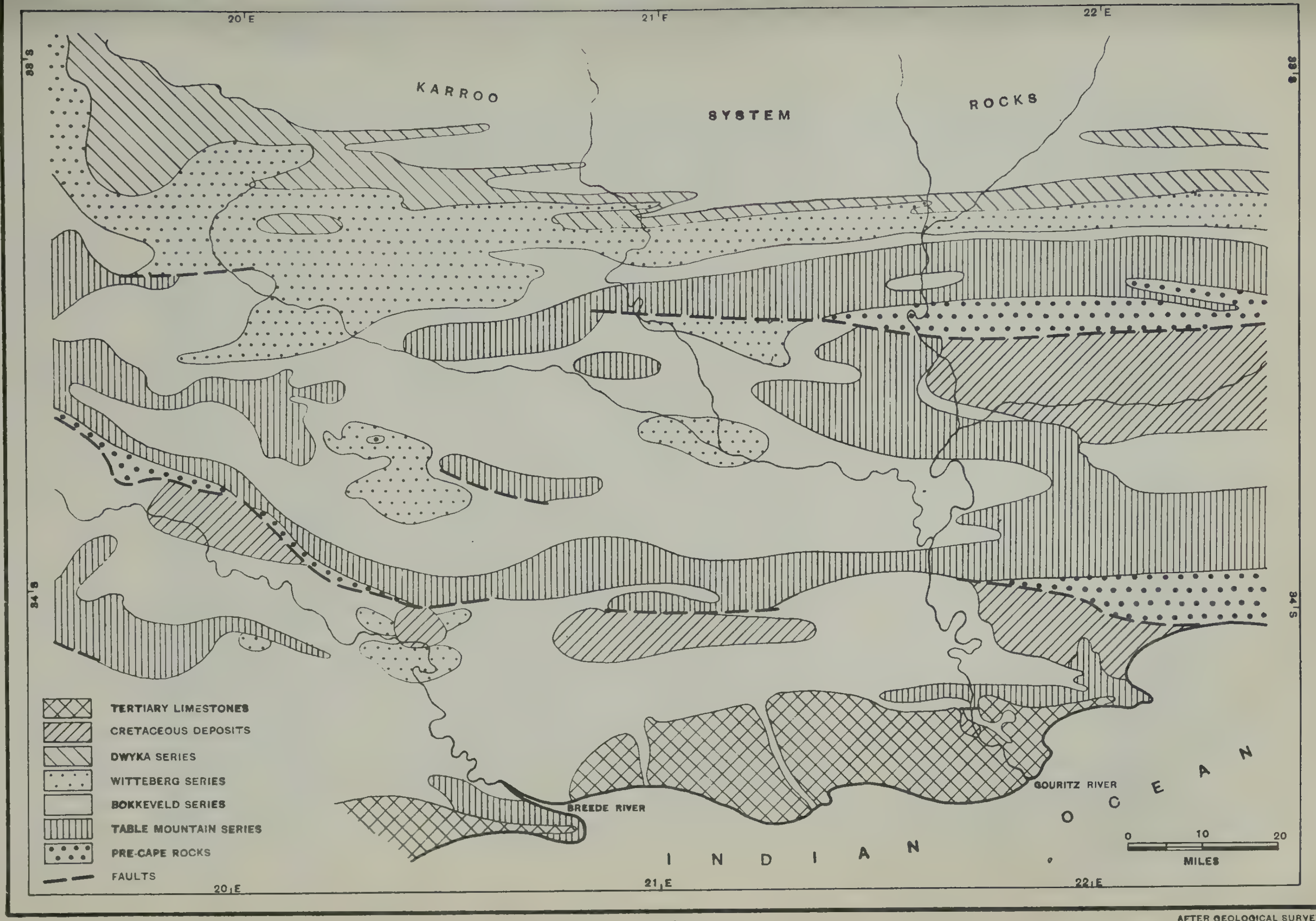

Figure 1. Geological Map

IV PRESENT-DAY CLIMATE, VEGETATION AND AGRICULTURE

The area may best be described as semi-arid although this is not strictly true. The temperatures in the lowland are never very low—slight frosts are sometimes experienced—but have a large dinural variation and high maximum (110°F.) In winter the mountain tops are often covered with snow, adding beauty to their grandeur. The rainfall varies from less than 5 inches in the open country to over 100 inches in the west against the mountains.

The effect of these climatic variations is seen in the vegetation. This includes the typical Karroo succulents and drought-resistant scrub in the open, while the kloofs in the mountains are well wooded. Aloes make vivid splashes of colour in the autumn.

Agriculture is confined to the narrow strips along the mountains and rivers where a supply of water is assured. Crops include grapes, tobacco, wheat, fruit, lucerne, etc. Sheep, goats and ostriches roam over the uncultivated parts.

V DRAINAGE

The area is drained by the Gouritz River System, all the major rivers of which are, for the most part, ephemeral; only the springs along the mountain fronts flow throughout the year. These streams derive their water from an accumulation of rain, mist and snow and are used for irrigation.

As a certain amount of confusion has been experienced regarding the names of rivers and poorts, the following description has been based on maps published by the Irrigation Department and the Trigonometrical Survey.

The drainage of the area can be divided into those rivers which rise to the north of the Langeberg Range and drain directly through them to the coastal plain to the south, and those which rise to the north of the Swartberg Range and, after flowing through the mountain and being joined by others rising in the area, flow through the Langeberg Range.

Of the first type there are three examples (the areas they drain are given to indicate their relative sizes):—

The Keisies-Appelkooskloof-Kingna Rivers which join near the town of Montagu before passing through Kogman's Kloof, drain an area of 71 sq. miles.

The Tradouws River flows eastward to the town of Barrydale and then turns south through the pass bearing the same name. It drains 125 sq. miles.

The Vette River, flowing through Garcia's Pass, drains about 9 sq. miles north of the mountain.

The second type, which display an apparent disregard for the high mountain chains, include the Touws, Buffels, Dwyka, Gamka en Olifants Rivers.

The headwaters of the Touws River lie to the north-west of the town of the same name. After flowing south-eastwards into the Little Karroo, the river is joined by tributaries from both the north and the south before it ultimately joins the Groot River. The total area drained by this system is 3,018 sq. miles, of which 793 sq. miles is north of the Couga Mountains.

The Buffels River has its source at the base of the Great Escarpment (Klein Roggeveld Mountains) and after flowing past Laingsburg, traverses the Witteberg Mountains at Leeukloof Poort and after crossing the intermontane area, cuts through the Klein Swartberg in Buffels River Poort. The trend of the course of the river

within the mountain has been likened to an incised meander because for a distance of 1·4 miles it flows northward before turning and again flowing southwards through the mountain. From the mouth of the poort it is called the Groot River. Before it joins the Touws River, it is joined by the Knuy River which drains the Ladismith area. To its confluence with the Touws River, the area drained is 2,166 sq. miles of which 1,533 sq. miles is north of the Witteberg. After its confluence with the Touws River it keeps the name of Groot River, passes Vanwyksdorp from where it begins to twist and turn, to its confluence with the Gouritz River.

The Dwyka and Gamka Rivers also rise at the base of the Great Escarpment and meet just before passing through the Swartberg at Gamka Poort. On emerging from the mountain it, too, turns back on itself and flows northward for a short distance. Before it enters the Oudtshoorn Basin it is joined by the Huis River which drains Seven Weeks Poort. It then passes Calitzdorp and flows due south to Jagtberg where it is joined by the Olifants River and flows through the Rooiberg-Gamka Hills at Jagtberg Poort. (This poort has no name on the maps; the local inhabitants refer to it as "Olifantspoort" or "Badspoort". These names are unsatisfactory as the former may be confused with the well-known Transvaal poort and the latter is too specific: the "baths" being located at the mouth of the Olifants section of the poort.) On emerging from the mountain it is called the Gouritz River where, south of its confluence with the Groot River, it crosses the Langeberg Range at Gouritz Poort and after cutting the low Aasvoëlsberg, enters the sea a few miles to the west of the town of Mossel Bay. Up to Jagtberg Poort and excluding the Olifants basin, it drains an area of 4,865 sq. miles of which 4,308 sq. miles are north of the Swartberg.

The Olifants River has its headwaters to the north of the Great Swartberg as the Tarka River. Within the Oudtshoorn Basin it is joined by the Kamanassie River and then flows westward to the Jagtberg Poort. It drains an area of 3,760 sq. miles of which 406 sq. miles are north of the Swartberg.

In all, the catchment area behind the Gouritz Poort is just over 14,000 sq. miles of which a little over half is to the north of the Swartberg Range.

The pattern of the main drainage lines is simple. With the exceptions of the areas within a radius of about 15 miles north of Gouritz Poort, where incised (N + 1) cycle meanders occur and a dendritic pattern is noticeable, and the region to the north of the Swartberg Range, a marked trellis pattern is both expected and encountered. Where certain rivers cut through the mountains the pattern tends to become almost rectangular owing to adaptation to the structures of the rocks. To the north of the Swartberg Range the pattern is again dendritic with the tributaries showing decreasing signs of subsequence northwards.

The poorts which are a striking feature of the Fold Belt, will be described in more detail later.

VI PREVIOUS LITERATURE

Descriptions of the area, including the deposits under consideration, are to be found in the reports of the Cape of Good Hope Geological Commission, the Geological maps and in most text books.

Rogers (1903) in his paper "The Geological History of the Gouritz River System" traces the river history of the area. He shows the existence of longitudinal valleys prior to the deposition of the Cretaceous deposits (Uitenhage Series), and the improbability of an antecedent origin of the courses of the Gamka and other rivers in

these words: "This absence of all traces of Pre-Uitenhage transverse valleys in the Zwartebergen and Langebergen together with the direct evidence of longitudinal valleys, which were eroded relatively more deeply than the present valleys, affords strong evidence against the view that the Gamka and other rivers from the Karroo were antecedent to the mountain ranges; in other words the rivers did not cut those valleys into rising mountains" (p. 379). He visualised the valleys as being filled with deposits derived from these mountains and the superimposition of the river courses from this surface. "We must thus look upon the river system south of the Karroo as a superimposed one as regards the folded country between the Karroo and the ocean; that is, it worked its way down from the surface (of Uitenhage Series) upon which the main streams were consequent, to a highly diversified surface . . . so that the rivers which thus maintained their courses came to have no direct relations to the structure of the country over which they ran after removal of the unconformable Uitenhage rocks" (p. 280). The Buffels River is suggested as having flowed over Garcia's Pass and having since been captured by the Gouritz River (p. 382). The gravels and deposits, he concludes, "record a past stage of the past development of the river system, during which the rivers had, in many parts of their courses, approached the limit of downward erosion . . ." (p. 383) and it was from this surface that the meanders in the lower reaches of the river were inherited.

Schwartz (1904) noted that the surface cut by the rivers was not a gradual one but stepped, because "behind each mountain chain there was an enforced period of base-leveling".

In his explanation of Cape Sheet 5, Rogers (1925) deals with the problem of the river evolution in this area once again. He argues (p. 9-10) that the unexpected course of the rivers through the mountains can be explained by one of three alternatives:—the rivers were antecedent to the mountain ranges; they were due to headward erosion (or as he says, consequent "on an original dip slope bordering the ocean"); or they were superimposed from a covering formation on which they were consequent. He reiterates his argument of 1903 that antecedency could not be the explanation, owing to the lack of evidence of pre-Cretaceous valleys through the mountains which would be expected. He then postulates the superimposition of the river on two points, namely the assumption that the Cretaceous deposits buried the mountain and have since been removed, and the form of the river which is like "a very deeply entrenched meander, shaped like that of a river flowing over flat ground". He dates this meandering as Early Tertiary or Later Cretaceous.

He quotes Prof. W. M. Davis (1906, ii), who favours the idea of extension by headward erosion and capture, "especially during earlier stages of whatever cycles of erosion the region has witnessed, for then the incision of valleys takes place most rapidly". Rogers rejects this possibility because of the bearing of the Uitenhage deposits. which are not preserved in the area which Davis saw, on the history of the river system.

Remnants of two erosion cycles are noted (p. 12)—the truncation of the Klein Swartberg at about 4,000 feet and the terraces at 2,600 feet on the northern flank of the Witteberg.

The conclusions reached by Rogers have had a profound influence on the later writers. Du Toit, for instance, suggests that the inauguration of the north-south drainage was a result of the crossfolding of late Cretaceous times (1939, p. 511). King (1942 and 1951) not only accepts without argument the postulate that the Buffels River once flowed through Garcia's Pass (p. 64 and 74), but also retains the hypothesis of superimposition for the origin of the major river courses. His views are presumably

based directly on Rogers' conclusions as can be seen from the following quotation: "Thus a diversified pre-Cretaceous or early-Cretaceous topography seems to have been smothered until no more than the tops of the hills, if that, remained exposed" (1942, p. 309). Even after du Preez had shown that the Cretaceous beds were deposited in tectonic basins (the bulk of the Enon antedating the faulting and crossfolding,) and Taljaard had suggested that most of the poorts were due to headward erosion, King clings to the superimposed hypothesis. This can be seen from the following quotation: "A recrudescence of folding then tilted the Enon formations within the valley, and the huge faults which now border the southern faces of the main ranges came into being. Possibly before this time, when the valleys were filled with detritus, the rivers initiated those courses across the ranges by which they now escape from the Karroo, often through the narrowest of poorts (e.g. Meiring's Poort), to the sea" (1951, p. 310).

Du Preez's paper on "The Lithology, Structure and Mode of Deposition of the Cretaceous Deposits in the Oudtshoorn Area" (1944) throws more light on the earlier history of the area. He finds (p. 225) that prior to deposition the area must have had relatively steep slopes and high relief with a relatively low gradient surface in the valley. Differential depression then occurred (p. 226) which, although it did not delimitate the depositional basin, modified it considerably. Delimitation was accomplished by crossfolding and the basin was deepened by faulting, possibly starting at the same time as the folding although the former was shortlived (p. 232). In this structural and tectonic basin the Enon deposits collected. The date at which the faulting ceased, is indeterminable; there is proof, however, that subsequent movement of about 15 foot has occurred in some places. (Prof. Taljaard, personal communication).

Taljaard (1948, i, p. 13) shows that no part of the Gouritz-Gamka River System could have been superimposed. He visualises the headward extension of the lower Gouritz River which breached the low Outeniqua Range along joints, during upper Cretaceous times. The Gamka breach was caused by streamhead extension from the Cango fault-scarp face, while poorts such as Meiring's and Towerwater are "without doubt" incised along major joints.

In his book "A Glimpse of South Africa" (1949) he gives a detailed and accurate account of the rivers and landforms in the area together with their possible dates. He maintains that the Touws-Groot-Gouritz River is the oldest in the region and that it flowed from the highlands around the present town of that name, through Gouritz Poort to the south, during Cretaceous times (p. 136). He considers the Gamka River as a fault-scarp stream and here dates the poort as Mid- to Late-Cretaceous in age. The Traka River (Towerwaters Poort) is of similar age (p. 140), while the Buffels is later but earlier than Prins Poort (see fig., p. 130), which he dates as Late Tertiary (p. 128-129). The Kogmanskloof, Tradouws and Vette Rivers are also all headward extended along joint planes of younger age.

VII REMNANTS OF THE TERTIARY SURFACES

General. Before describing these remnants in particular, a few general characteristics may first be described. The flat-topped terrace remnants and the planed surfaces have been reflected in names such as Grootvlakte, Rooivlakte, Kareevlakte (all surfaces), Platrug, Platrand, Tafelberg, Long Mesa and Jakhalsplaat, while the local inhabitants refer to the terraces as "plate" (sheets) and "platos" (plateaux).

For the most part the terraces are confined to the flanks of the mountains, usually within 5 miles of the latter. They overlie with an angular unconformity older

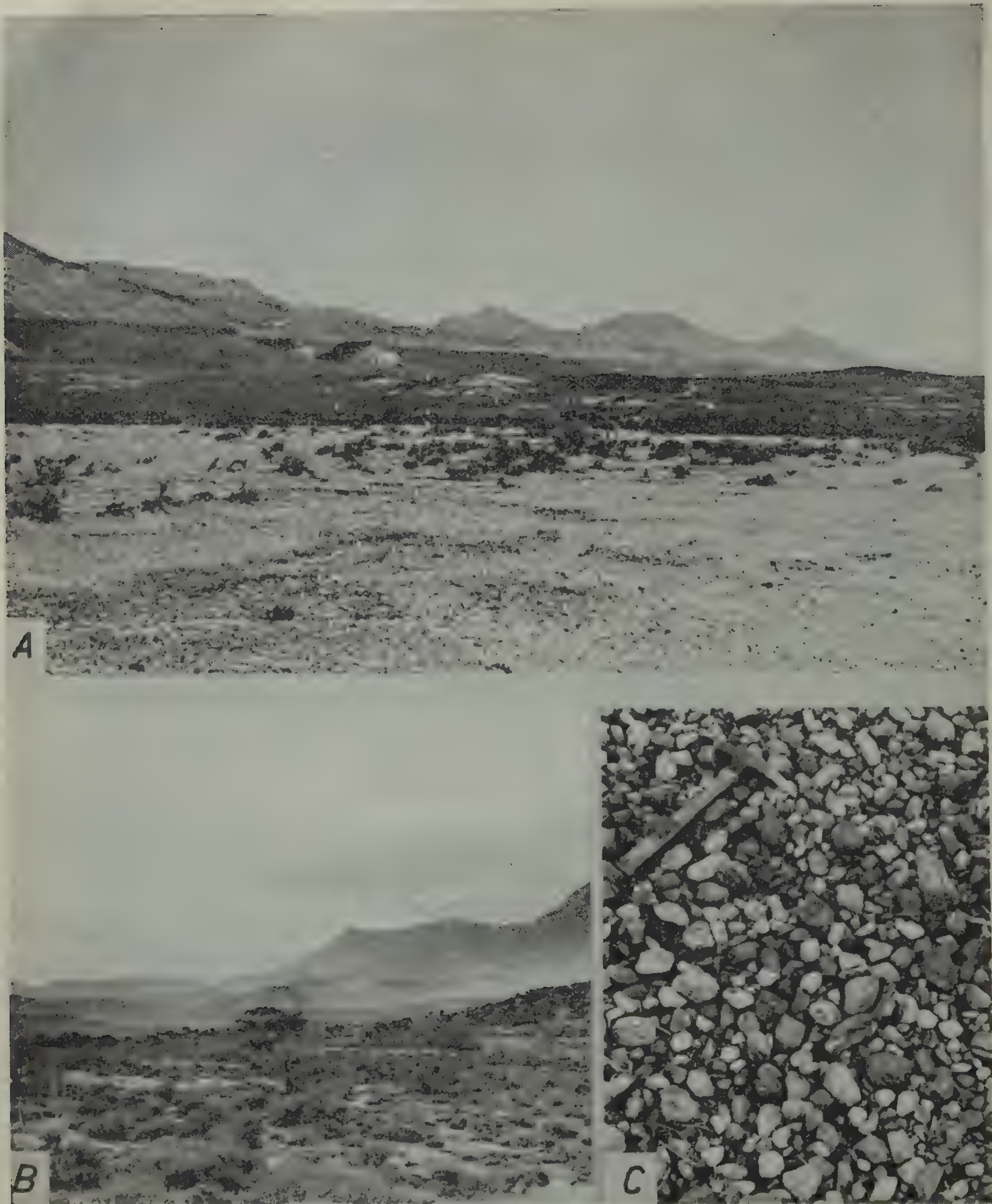

Plate I

(a) Consolidated gravels capping remnants of higher surface on northern flank of Langeberg, east of Garcia's Pass.

(b) Remnants in the Lemoenshoek area showing "stepped" upper surface on skyline, looking east.

(c) Pebble patch on a remnant's surface (Long Mesa).

rocks—in the western area those of the Cape System (usually Table Mountain and Bokkeveld Series, but where the faulting has brought the Witteberg Series close to the Table Mountain Series, they may overlie both); in the east, rocks of the younger Enon and older pre-Cape and granitic rocks; to the north, Witteberg, Dwyka and Ecca Series.

In some cases the terraces are contiguous with the mountains, but generally speaking those having consolidated deposits are isolated from them, and a bench, at a height continuous with that of the surface underlying the deposits, is often discernible on the flank of the mountains. However, both on low and high dipping strata the jointing and bedding planes produce a number of similar "benches" at various heights, and thus the use of this feature has been restricted to those which are in all probability a continuation of the surface. The terraces are usually diamond-shaped, having a maximum width some distance from the mountain.

The terraces occur in groups and any deposits resting on the surfaces, although varying considerably from place to place, are usually similar in neighbouring terraces. The thickest deposit measured (including recent scree) to the north of the Touwsberg, were 255 feet—the thickness increasing towards the mountain. Various degrees of cementation of the deposits have taken place, generally speaking confined to the lower 50 to 70 feet. Where the deposits overlie shales it is common for the latter to have been leached to clays of various colours.

As will be shown later, the surface is nearly everywhere overlain by gravels which in some cases have been cemented to form a conglomerate; the gravel consists of rounded boulders, cobbles, pebbles and granules (the size being smaller and the rounding better towards the valley, and regionally towards the south-east).

The Northern Slopes of the Langeberg. (S ee Fig. 2). The consolidated deposits stretch from Barrydale in the west to Gouritz Poort in the east. They occur in five groups: those in the vicinity of Lemoenshoek; Brand River; Long Mesa (including those to the west of Garcia's Pass); Sandkraal and those to the north of Gouritz Poort, but they are so similar that they will be described collectively. Their preservation must be attributed to their lithology which includes con glomerates and surface quartzites (silcretes). These terraces are often bounded by cliffs to the north but the southern ends are invariably covered with slump material.

Figure 2.

Consolidated (solid) and unconsolidated (open) Gravels on the Northern flank of the Langeberg. From Lemoenshoek to Gouritz Poort.

The height of the surface underlying these deposits was measured in a number of places and from this the following calculations were made.

(i) Relative to the streams flowing near them, they show incision depths varying from 300 feet to 850 feet. Readings obtained at the northern edges of the terraces were as follows:

Gouritz Poort	...	...	872 feet
Sandkraal	...	...	305 feet
Garcia's Pass	...	...	300 feet
Lemoenshoek	...	...	515 feet

(ii) The surface appears to be concave in a valleyward direction. There is nearly always a sharp angle where it abuts against the mountain. The results of the measurements in this direction are:

Lemoenshoek. A fall of 110 feet over 0·75 mile or 146 feet per mile near mountain. followed valleywards by 135 feet over 1·00 mile or 135 feet per mile.

Long Mesa. A fall of 300 feet over 1·25 mile or 240 feet per mile near mountain.
followed by 60 feet over 0·5 mile or 120 feet per mile.
followed by 25 feet over 1·5 mile or 16 feet per mile.

Sandkraal. A fall of 75 feet over 0·25 mile or 300 feet per mile near mountain
followed by 80 feet over 0·5 mile or 160 feet per mile.

Waterval (S. of Sandkraal) 85 feet over 1·0 mile or 85 feet per mile
followed by 25 feet over 0·25 mile or 50 feet per mile
followed by 80 feet over 1·25 mile or 53 feet per mile
followed by 30 feet over 0·25 mile or 60 feet per mile.

(iii) The average of these 4 profiles is 106 feet per mile.

(iv) The southern edges of the remnants (or most southerly measurement in some cases) show a regional inclination to the east. In places, notably Muiskraal (north-west of Garcia's Pass) and Tafelberg, a slight back slope was measured amounting to 15′ and 10′ respectively. (These figures serve to indicate a measurable back slope rather than an accurate measurement.) The regional slope from west to east can be seen from the following list of heights. The reason why they appear irregular is that they are not all against the mountain.

Witkleikop (Lemoenshoek)			2,265	feet
Brand River	...	...	1,985	,,
Long Mesa ...	...	...	1,872	,,
Muiskraal ...	...	...	1,860	,,
Sandkraal ...	...	...	1,655	,,
Waterval ...	...	...	1,534	,,
Poort ...	...	...	1,200?	,,

(v) The direct distance between the first and last readings above is about 50 miles and thus the average slopes is 20 feet/mile; along present-day drainage lines it is about 65 miles or 16 feet/mile.

Although the slope of the surface remnants is greater to the north than to the east, they must have flattened considerably valleyward, as the east-west striking ridges of Bokkeveld and Witteberg Series appear to have been planed to a common level. Some places show that they must have been higher than the surface; usually this occurs near the mountains and separates the groups. The best example is that between Lemoenshoek and Brand River where the beacons at heights 2,048 feet and 2,079 feet

are well above the projected level of the surface. Similar phenomena can be seen between the groups in this and other regions.

The tops of the terraces are usually smooth and possess a valleyward slope of low inclination in the north, increasing gradually to the south. Some, however, appear stepped, that is, they have two surfaces at different heights as at Lemoenshoek. This can be seen from the terrace on the skyline in the photograph. (Plate I, B.)

Patches of small rounded pebbles occur on top of nearly all the remnant terraces, these being exceptionally well displayed on the largest (2·97 sq. miles) on which the beacon Long Mesa stands. They vary from 3 inches to less than 0·25 inches in length, and although all sizes occur, one average size of one patch usually differs from the next. They are purple, black or reddish-brown in colour interspersed with white quartzite and sandstone pebbles. Some of those of darker colour are undoubtedly weathered quartzite and sandstone, but that some have been derived from the ferricrete, cannot be ruled out.

The consolidated deposits resting on the surface show a certain similarity in succession, even over the large area, as can be seen from the table of sections in Fig. 3. The general succession they suggest is that the surface is nearly everywhere immediately overlain by a well-rounded pebble layer of varying thickness which is partially cemented and weathers easily. The pebbles are less than 1 foot in length and average between 3 inches and 6 inches in length. This layer grades upward into cobbles and boulders which are better cemented by silica and iron oxide. Above this, often without a sharp contact, rest the silcreted sands and pebble lenses (conglomerate). The pebbles are all well rounded and average a few inches in length. One fairly persistent band of conglomerate about 4 feet thick can be seen in the Sandkraal area. This is usually overlain by less well consolidated deposits—often a pebble or cobble ferricrete above which sand, soil and rubble derived from the mountains have accumulated. This scree may be over 100 feet thick near the mountain, but valleywards it is considerably less, if present at all.

Two interesting features of the succession near Derde River Beacon are that the deposits contain a sandstone horizon which has a white clay-like matrix and that this horizon also shows signs of cross-bedding.

Although a few of the remnant surfaces have no deposits other than a thin veneer of gravels, they are easily recognizable as they overlie dark shales and the gravels are composed of light-coloured quartzites and sandstones. Little attention was paid to these deposits, with the exception of the area to the north of Gouritz Poort. They lie at a lower level than the consolidated deposits in most cases, although in the Long Mesa area they appear to be the valleyward extension of the top of the terrace. Their distribution is along the present drainage lines and in the valley they are about 200 feet to 300 feet below the planed ridges. The occurrences of loose unconsolidated gravels in the Gouritz Poort area have already been mentioned. They occur on both the east and west banks of the river. Those on the left bank occupy a large flat expanse which has rounded edges covered with well rounded and chatter-marked boulders and pebbles. The height of the top of this plain is between 1,000 and 1,100 feet, and can be seen to be related to the change of slope shown by the poort (see later). On the right bank of the river, below the trigonometrical beacon at a height of 759 feet, 20 to 25 feet of gravel is seen. For the most part this is unconsolidated, showing a slight ferruginous cementation in the lower parts. The components of this gravel are all well rounded, and some of the larger boulders are extremely well chatter-marked on all sides.

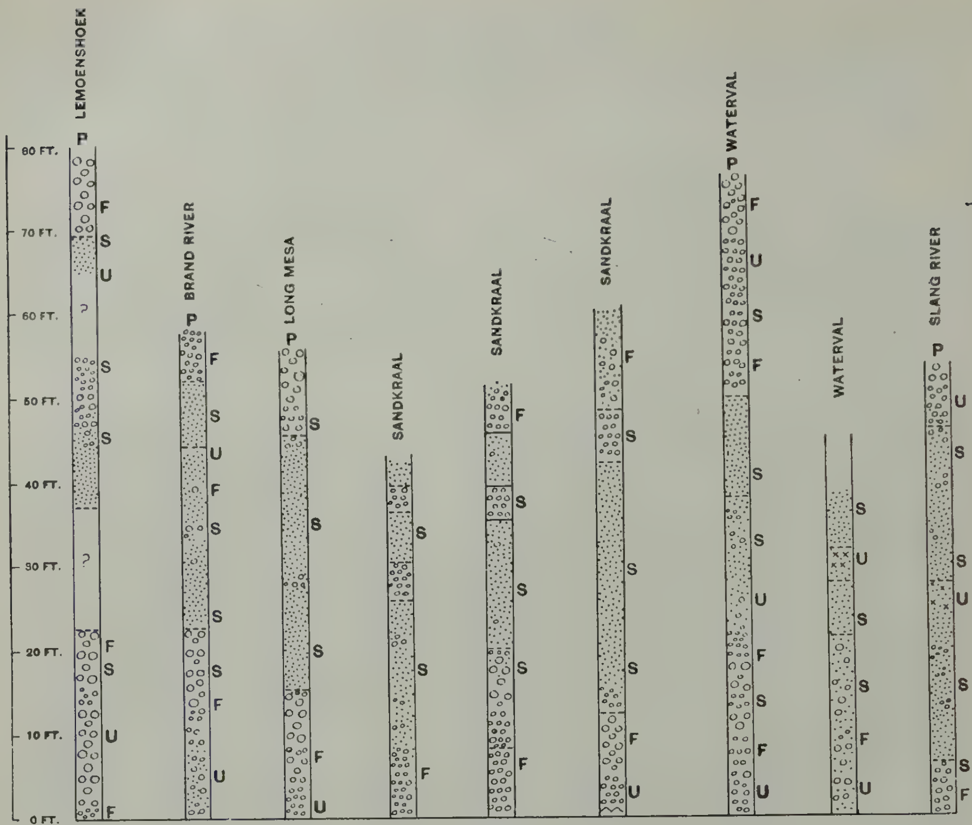

TABLE OF SECTIONS

Figure 3

Dots indicate sand
Small circles indicate pebbles
Large circles indicate boulders

F denotes mainly ferruginous cement
S denotes mainly siliceous cement
U unconsolidated
P denotes pebbles present on upper surface

The Warmwatersberg Area. On the southern side of this mountain a number of terraces occur at heights of about 1,750 feet. They are covered with a loose, unconsolidated layer of gravel. On the northern side two terraces were noted at heights around 2,000 feet, both covered with a thin veneer of gravels from 15 to 25 feet thick. They overlie Bokkeveld Series beds and the surface is remarkably smooth, being cut across the sandstones and shales alike. The sizes of the components of the gravel range from sand particles to 2-feet boulders near the base and up to 5 feet near the top. They are angular to sub-angular in shape and mainly composed of Table Mountain Series quartzites and sandstones.

To the south-east of Bellair Dam there are a few sporadic patches of gravel. These appear to have been reworked. To the south of this dam there is what appears to be a perched, disused narrow poort through the mountain. It is locally known as the "Valley of Desolation", and notwithstanding the legendary curse of the original Hottentot inhabitants, a more probable reason for the lack of vegetation and failure of crops is suggested by the fact that a borehole passed through 165 feet of sand and boulders without reaching bedrock.

The North-west. In the north-western portion of the Little Karroo, bounded by the Waboomsberge, Coega Mountains and the Anys-Witteberg Range, a large tract of the land has been planed to a smooth, though gently undulating surface. (See Plate II, A). This could be inferred from the name of the farm and the beacon in the area, namely Grootvlakte. The general height of this surface is about 2,500 feet. It rises towards the mountains and is covered in places by a thin sheet of gravel and soil. It slopes towards the Touws River from both north and south, and in the north the planing of hard and soft rocks is beautifully displayed (see Plate II, B). It can also be seen from the photograph that a lower surface exists at about 100 feet below the higher surface, below which the river has been entrenched.

Touws River Town Area. The country to the north-west of the Couga Mountains, to the west of the town of Touws River, is again very flat, at a height varying between 2,500 and 3,000 feet. Taljaard (1948, p. 77) has described relics of an older surface which cuts across rocks in the area about 1,000 feet higher than the above-mentioned surface, at heights of between 3,600 and 3,800 feet, reaching 4,000 feet at a maximum. This corresponds in height to the remnants of a surface mentioned by Rogers (1925, p. 12): "During the earlier of the two cycles there was produced the flat surface which truncates the anticlinal fold of the Klein Zwartberg". "The height of the remnants of the earlier formed plain is about 4,000 feet".

Anysberg. On the southern slopes of the Anysberg both the surfaces mentioned above are displayed. The lower surface appears in certain places to merge into the higher, and the deposits on the former consist of unconsolidated or weakly-consolidated rounded gravels with boulders of up to 3 feet, at a distance of 3 miles from the mountain. The slope of this surface was not measured, but it can be seen sloping towards the present river from the north, while in the south the present-day surface also slopes in the same direction, as at Karreevlakte.

The deposits of the higher surface are invariably cemented by iron oxides with some silica, the relative proportion of Fe to Si being 3 : 1 at Wilge River. This cements boulders and cobbles of Table Mountain Series and vein quartz.

The remnants to the south of Touwsberg are essentially similar, there being possibly a smaller proportion of consolidated gravels, and their general height is naturally lower (approximately 2,100 feet for the higher surface).

Touwsberg-Swartberg Intermontane Region. (Fig. 4.) The area to the north of the Touwsberg is more complex as regards the surfaces; the writer noting no less than five surfaces possibly all different, above the present river alluvium. However, owing to the position of the mountain in relation to the local erosion base, it is probable that these are contemporary surfaces which have been formed at different heights. Fortunately the highest two surfaces can be recognized fairly easily.

Midway between the ends of Touwsberg, and sloping away to the north, the higher surface is represented by a terrace which overlies T.M.S. and the Bokkeveld

Series to the second shale horizon. It carries an extremely thick covering of deposits, the base of which is 650 feet above the local stream and at an altitude of about 2,450 feet. The contact of this surface with the mountain appears to be a sharp one.

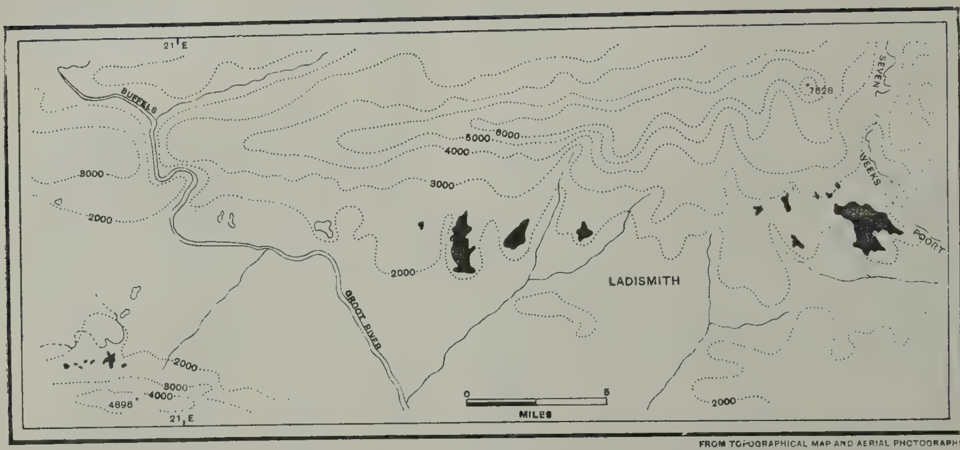

Figure 4

Higher (solid) and lower (open) surfaces on the Southern flank of the Klein Swartberg. From Buffels River Poort to Seven Weeks Poort. Touwsberg in south-west corner; Amalienstein in the east.

Resting on this surface is a layer of ill-sorted, more or less consolidated layer of pebbles, cobbles and boulders, between 50 and 80 feet thick. On the south-eastern edge of the terrace it is predominantly cemented by iron which weathers easily, forming caves and overhanging ledges, whereas to the north and west the silicious conglomerate is again encountered . . . this forms a hard cap and cliffs of about 30 to 40 feet high, below which blocks of conglomerate are littered over the hill slope. The jointing in this area passes both through and around the pebbles, depending on the degree of cementation. The size of the boulders in the cemented portion seldom exceeds 5 feet in length, but some of as much as 10 feet were seen near the mountain.

Above these deposits there is an accumulation of scree and soil containing huge pieces of angular quartzite up to 20 feet in length. The thickness measured was approximately 200 feet, but this must be exceeded further southward where debris is still accumulating.

Similar terraces, though none with the thick scree deposit, are found to the west of this terrace. To the east there are no remnants or even reliable benches. The surface dips to the north-west, the valleyward inclination near the mountain is 125 feet/mile and the westerly component is about 80 feet/mile.

To the north of the central terrace and about 100 feet below it, a second more extensive surface can be seen which is separated from the former terrace by about 75 yards. It stretches to the north and north-west at a general height of 2,000 feet (from 2,243 feet to 1,903 feet) with a slope averaging 120 feet/mile over 3 miles. It is covered with a gravel mantle in places, 10 to 20 feet thick and composed of well rounded rock fragments and soil, the former averaging a few inches, but also reaching up to a few

Plate II

(a) The surface near Grootvlakte looking N.E. (Anysberg left background).

(b) A bevelled remnant in the Bloutoring area with the Touws River in the foreground. Anysberg on the right. Taken from lower surface.

(c) Stream-cut bench on west side of Jagtberg looking S.W. over incised Gamka River. Oblique to strike.

(d) Structural bench on south side of Jagtberg. Confluence with Gamka River middle left. Olifants River incised on left.

feet in length. At a distance of 3 miles from the mountain they still attained a length of 2 to 2·5 feet.

In places a third surface is seen cutting this in the south, but they merge into one another in the north. A fourth can be seen still lower down.

On the farm Mierefontein there is a bench which is about the same height as a small remnant terrace to the east of the Ladismith-Laingsburg road, where it begins to ascend the mountain. It is situated on the watershed of the Groot and Touws Rivers at an altitude of 2,026 feet, and carries a layer of about 10 feet of unconsolidated gravels. To the east 120 feet below this, a smaller remnant occurs.

At a point due north of Touwsberg the Groot River emerges from the Buffels River Poort. Just to the east of the poort lies the first of a number of terrace remnants which stretch to Voorbaat at an altitude of about 1,700 feet—350 feet above the local river height. They have an easterly inclination of about 20 feet/mile and are covered with unconsolidated gravels.

The Dwars River (Ladismith) Area. To the east of these terraces and 675 feet higher, the higher surface remnants are seen on the southern flanks of the Swartberg at an altitude of approximately 2,400 feet on top. The surface on which the deposits rest is lower than this, and the lowest 50 feet or so are consolidated into conglomerates. The top of the deposits is remarkably smooth and covered with small rounded pebbles as described on the southern terraces. The soil and rainfall here is such that the vegetation is thicker than on other terraces. The surface on which the deposits rest shows an easterly slope of about 17 feet/mile.

Amalienstein. To the east of Ladismith, in the vicinity of Seven Week's Poort, there are a number of remnants, the largest of which covers an area of 2·21 sq. miles and is situated just to the west of the poort. This impressive terrace has a covering of about 35 feet of conglomerates overlain by 40 to 50 feet of scree, at the northern end. The conglomerate is again a mixture of sizes from boulders to sand which have been cemented by silica and iron oxides and the base is at about 2,806 feet. No remnant straddles the fault. The terraces have a westerly slope of 10 feet/mile. To the east a pronounced level occurs, cutting the T.M.S. at a height of about 3,000 feet. On the ridges in the valley reworked gravel occurs at about 2,000 feet which might be a reflection of the lower surfaces seen elsewhere.

Rooiberg. To the north of Rooiberg a few terraces were seen but not visited by the writer. From the topographic map they appear to be at heights of between 2,000 feet and 2,500 feet. The western end of the mountain has a number of benches at a height of about 2,000 feet, while the area to the south has a number of terraces covered with loose unconsolidated deposits at an altitude around 1,400 feet. There are no conspicuous terraces above these surface remnants, although it must be stated that this mountain (a simple anticline) is stepped by "benches".

The South-east. Only in the valley drained by the Slang River, between the Pogha and Gamka Hills, are consolidated gravels again found (Fig. 5). They are extensive and well preserved on the southern side of the valley where they overlie both T.M.S. and Bokkeveld Series, while on the northern slopes they generally overlie only the Table Mountain Series. They rise from heights of around 1,500 feet to well over 1,800 feet to the east. A remarkable fact of the deposits in this area is that they are less well cemented and show definite signs of crossbedding in the sandstones. A general section (a combination of three) shows the following succession (see Fig. 3). The surface

is overlain by ferricrete or conglomerate of small pebbles which grade upward into cobbles and boulders. This varies from about 5 to 10 feet in thickness. On this, 8 to 20 feet of silcrete, sandstone with conglomerate and pebble lenses occur which in turn grade into the usual gravel and scree on the top. The tops of the remnants are flat and often covered with pebble patches.

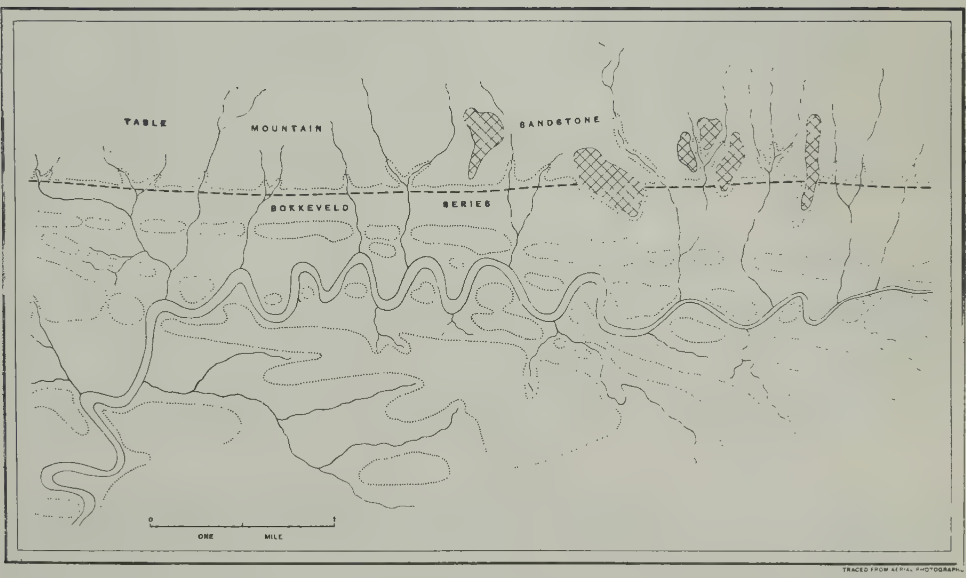

Figure 5

The Slang River Valley showing approximate Table Mountain and Bokkeveld Series contact (broken line), Eo-Oligocene consolidated gravels (cross-hatched) and more resistant ridges of Bokkeveld Series (dotted)

Terraces were also seen in the narrow valley of the Kamma River, a few miles south of the Slang River, and here too the remnants are confined mainly to the southern slopes.

The Oudtshoorn Basin. Although no measurements were taken in the Oudtshoorn basin itself, one or two observations must be included here. The "High Level Gravels" have been described by a number of authors. Du Preez (1944) refers to a lower surface sloping towards the Olifants River and cutting across the Enon deposits. In the Kamanassie Valley to the east, de Villiers (1938) finds that the two surfaces are 200 feet apart, the lower about 400 feet above the master stream. The lower surface is preserved on the southern side of the river only, while the higher has a maximum distribution on the northern side at an altitude of about 2,100 feet (topographic map). If this higher surface is correlated with the lone remnant a few miles to the north-west of Jagtberg Poort a regional slope of about 10 feet/mile is obtained. This latter remnant is at a height of about 1,760 feet.

In this western portion of the Oudtshoorn basin there are far fewer remnants than in the central and eastern portions. However, those at Armoed and Kerkrand

dip towards the centre of the basin and not to the present more southerly position of the Olifants River. Benches are, however, seen everywhere against the mountains, occurring at various altitudes, none bearing gravels. One soil-covered bench is of special interest: it is cut into the west side of Jagtberg, parallel with the Gamka River at an altitude of just over 1,200 feet at a slight angle to the dip of the T.M.S. (Plate II, C). It stretches for a distance of about 0·75 mile from the T.M.S.-Bokkeveld contact to the south. Two chatter-marked boulders and a small piece of calcrete, containing rounded pebbles and embedded in the soil covering, provided the only positive evidence that the bench was formed by water action. This bench today slopes towards the present river to the west and to the south, and is the only water-cut bench which could be found. That on the southern side, parallel to the Olifants River, is a structural bench (Plate II, D), and the general height of the shoulders and benches on the southern side of the river suggests a slope to the east. No higher benches were seen.

Swartberg-Witteberg Intermontane Region. In the whole of the intermontane area between the Swartberg and the Witteberg only a few patches of unconsolidated gravels were seen, near the Buffels River and the Klein Swartberg River. They occur at a height of just over 2,000 feet. An investigation of aerial photographs showed no other occurrences. Considering present-day stream activity to the north and south of the Klein Swartberg, the absence of the higher surface cannot be ascribed to more active weathering and stream-cutting in the intermontane region. In fact, the reverse is true as, owing to the higher precipitation, the streams on the southern flank are more active.

The Area South of Laingsburg. North of the Witteberg a number of terraces occur. These have been described and studied by both South African and overseas geologists.

No less than three surfaces were seen in this area (Fig. 6) and Strydom (1950, p. 268) mentions "two or three" river terraces below these. The most extensive surface is found on both sides of Leeukloof Poort, sloping towards the poort (80 feet/mile from the east) at an altitude of about 2,550 feet, but it has larger remnants on the western side where the largest terrace covers an area of 4·64 sq. miles. This huge terrace stretches for about 4 miles northward from the mountain and at its broadest point is about 3 miles wide. Near the mountain it both grades into and abuts, with a 130-foot scarp, against a higher surface. Where the two grade into one another the slope is more than 300 feet/mile, while the northward gradient is merely 140 feet/mile. Where the two surfaces are separated by a scarp, they are situated to the south of a ridge of Dwyka tillite which has been planed down to the level of the terrace to the west. The surfaces grade into one another to the west of this point. The width and slopes of the gap, however, suggest that it was in all probability in existence when the upper surface was formed, as the latter also slopes in this direction (10 feet in 200 feet).

The upper surface can also be seen further to the west, where it is preserved behind more resistant ridges of the Witteberg Series quartzite.

The ridges composed of the Karroo System rocks to the north all present the appearance that they have been planed to a common level at just over 3,000 feet, with some resistant parts protruding above this.

A stream has cut into a large remnant at a distance of about two miles from the mountain, exposing the shales and tillites of the Dwyka Series. Here it has cut a definite bench about 25 feet below the higher surface. In a downstream direction the difference between the two surfaces increases to about 75 feet. A similar feature may be seen to the east of this point.

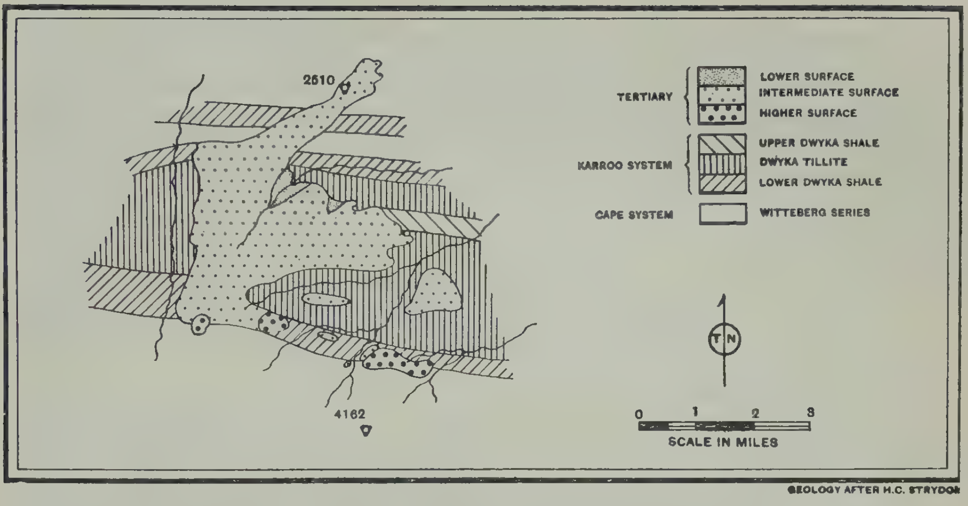

Figure 6

Sketch Map showing Geology and surfaces on Northern flank of the Witteberg, south of Laingsburg.

The components of the unconsolidated layer of gravels overlying the large surface are remarkably constant, varying from boulders to sand, the size of the largest boulders being about 10 feet near the mountain but decreasing to about 3 feet within the first mile and virtually remaining this size right up to the northern end of the terrace. The average size, however, decreases considerably to a foot or less. Throughout the larger boulders have percussion marks, and the smaller ones show some degree of rounding. They are undoubtedly Witteberg quartzite derived from the mountain to the south. The northern edge of the terrace was measured to be 355 feet above the Buffels River.

Terraces were also noted on the watershed between Hartebeestspruit and Kerk River to the east of this group, i.e. between the Doekberg and the Elandsberg, at about 3,000 feet (topographical map). They appear to be covered with unconsolidated gravels.

Die Hel. The intermontane valley in the Swartberg, known as 'die Hel,' also shows from a stereoscopic investigation terraces on the southern side of the valley, and the Bokkeveld ridges appear to have been planed down to about the same level as this surface.

Other Drainage Basins. The only other areas of importance are those which are drained by short rivers rising north of the Langeberg and flowing directly out of the Little Karroo. In these areas the only gravel patch which was found was on the Droogte River in the Kogmanskloof drainage basin, a few feet above the local stream and about 50 feet above the master stream.

On the watershed between the Tradouws and Kingna Rivers a few loose T.M.S. boulders were seen. These were lower than shoulders and benches seen all along the mountain range to the south, which suggests the former existence of a surface rising to about 3,000 feet in the Montagu region.

The Area South of Gouritz Poort. This area was visited briefly to compare the nature and heights of the Tertiary "Silcretes, Ferricrete, and Older Gravels" described in the Explanation to Sheet 201 (Haughton, etc., 1937 (i)) with those north of the Langeberg. They occur at about 1,100 feet both to the east and west of the poort. They are inclined towards the present river and drop to just under 500 feet north of Aasvoëlberg, an average slope of 38 feet/mile. It must be remembered that the warp axis to the south of Riversdale may extend and influence this figure (see later).

VIII PETROLOGY

A certain number of specimens were collected in the field from various localities and horizons, but the types differ so little both in hand specimen and in thin section, that they will be described collectively.

Ferricrete. Samples from Wilge River (near Anysberg Post Office), Touwsberg, and from the area to the north of the Langeberg were examined. They consist of T.M.S. pebbles, usually well rounded, together with quartz grains which have been cemented by a dominantly ferruginous material, giving them a deep red or brown colour. On weathering the matrix is usually removed first, sometimes resulting in a pitted or "honeycomb" effect as in the Slang River valley.

Calcrete. Only one sample of this was found on the bench at Jagtberg. It contained a few well rounded pebbles and quartz grains cemented by lime carbonate.

Silcrete. This type is confined to the areas to the south and east, samples being collected at Lemoenshoek (not pure), Long Mesa, Sandkraal and Tafelberg. It is a greyish-white rock, often streaked with brown staining and containing small cavities. It usually contains a number of small, well rounded pebbles, but in its more pure form, as seen at Tafelberg, is composed of subangular, angular and slightly rounded sand-sized grains of quartz. (Wentworth scale is used throughout as the rocks were studied both in hand specimen and thin section.) Its fracture is sub-conchoidal to splintery and it weathers along cracks or exfoliates. When struck with a hammer it produces a metallic "ring". In thin section it is seen to consist mainly of granules and sand particles composed of quartz and cemented by chalcedony. Zircons are the most abundant of the accessory minerals. They are rounded but their size and abundance appear to vary unpredictably.

Other accessory minerals that were noted, were muscovite, tourmaline, hematite and limonite. The quartz grains are separated from one another and nearly all show signs of strain (undulatory and columnar extinction). The matrix may comprise as much as 65% by volume, but is usually between 40% and 50%. Vague colloform structures were observed, as were differences in the amount of iron staining. A few of the grains have fretted outlines.

Conglomerate. There is a continuous transition between this type of rock and a silcrete containing pebbles. The conglomerate here may be said to consist of cobbles, pebbles, granules and grains (in some cases even boulders) of T.M.S. quartzite and vein quartz set in a matrix of silica and iron oxides. Excluding the particles larger than granules, the ratio of cement to allogenic material, is about 1 : 1, although this proportion is not by any means constant. Usually the cement is such that the con-

glomerate breaks through pebbles and matrix alike, but where the cementation is weaker, it breaks round the pebbles. In places abundant colloform structures and rhythmic layering were seen, varying in size from 0·6 cm. to 0·2 cm. (see Fig. 7), presumably where silicious and ferruginous material have been coagulated together. (See Frankel and Kent, 1937). A peculiarity of some of the conglomerates is the occurrence of what appear to be silcrete "pebbles". These have a much higher iron staining which increases towards the edge of the pebble, but the proportion of grains to matrix is lower than the purest silcrete found. (See Fig. 7.) In the conglomerates some of the grains of quartz have fretted outlines. Zircons and other accessory minerals are also present in small quantities in the matrix. In places the conglomerate developes a jointing normal to the bedding and at the edges of the terraces large blocks can be seen lying on the slopes below the deposits.

Figure 7

A. Silcrete (?) Pebble and Rhythmic Layering in Conglomerate.
B. Colloform Structures in Conglomerate.

IX POORTS IN GENERAL

Because of the significance attached to the gorges or "water gaps" in determining the history of this area, special attention will be paid to these striking and beautiful features of the Cape Fold Belt. A poort is a relatively narrow, steep-walled

opening cutting almost perpendicularly across a topographic barrier, through which the more open areas on either side are connected, usually by a river.

Five major types can be recognised, namely, consequent, antecedent, subsequent, inherited and superimposed poorts.

A consequent poort is one which has developed in response to the inequalities of a newly formed surface. They may develop on coastal plains for instance (see von Engeln, 1942, p. 121).

An antecedent poort is one which results from an uninterrupted flow of a river, across which the barrier is formed. A necessary but not essential condition for this type of poort is, that the age of the structure succeeds that of the establishment of the river.

A subsequent poort is one which is developed along a relatively weaker part of the barrier by a headward eroding stream.

An inherited poort is one in which the stream responsible for the formation of the poort has been imposed from a higher level having the same governing structure.

A superimposed poort may be considered a special type of inherited poort, as has been shown to be the case with rivers (Maske, 1951). It is a poort which owes its existence to the letting down of a stream from a foreign structure with which it was in harmony, the stream maintaining that course in the lower structure. A necessary, but again not sufficient, condition for this type of poort is, that the downcutting stream must pass through an unconformity.

For three of the above types poort formation may take place when a river has reached any stage of development; but for subsequent and consequent river-poorts youth is essential.

It is impossible, for obvious reasons, to draw up criteria for the recognition of the different types of river-poorts listed above. The type can only be decided after taking into account the history of the area, the physiographic age of the area and rivers and the form of the poort.

However, subsequent poorts in similar climatic regimes, traversing similar geological formations and structures, and in regions of physiographical stages of development similar to those displayed in this area, make an interesting study. The relationship between their age, degree of adaptation to structure, valley slopes and trend of their courses will be traced in general, drawing on examples displayed in the area.

As this type of poort has its origin in the headward erosion of a stream, it is characteristically controlled by the structure and lithology of the rocks.

In youth the poort is a narrow twisting defile with the river channel parallel to the directions of the major joint and bedding planes. The walls of the poort are steep and often vertical. One of the many examples of subsequent poorts of this age is that of Seven Weeks Poort. Some idea of the winding pattern of the river course can be seen from Fig. 4. The total relief can be pictured when it is realized that the highest point of the Swartberg Range in this area (7,628 feet) lies 2·5 miles to the west of the poort and the river flows at a level of around 3,000 feet. Kogmanskloof, Tradouws Pass and Prins Poort are all examples of similar origin and age in the area.

As maturity is approached the river tends to straighten its course as a result of the increased gradient through the poort and a lessening in the degree of structural adaptation. Side streams reduce the precipitous slopes of the walls to a smoother gradient. Control by the minor structural features is lost, although the general trend of the course may still reflect the major structural lines. This stage of development is

displayed by such poorts as Gamka, Jagtberg and Buffels Poorts. That of late-maturity, by Gouritz Poort.

Old age may be presumed to have been attained when the river's course is not influenced by the structural features and the walls of the poort have become low in gradient and comparable to those of the mountains through which it passes.

It must be emphasised that nearly all poorts in this area have been inherited, and as a result of the rejuvenation youthful features are nearly always seen.

X RECONSTRUCTION OF THE DRAINAGE LINES

As has been previously mentioned, the key to the river evolution lies in dating the breaching of the mountains. Associated with this is the problem of the origin of the poorts. With the first point in mind, an attempt will be made to reconstruct the drainage lines by correlating the various surfaces, the latter on grounds of lithology, their altitudes and their mutual relationship. Possible dates of the surfaces will be discussed later.

Setting a date to the Gouritz Poort is relatively easy, as the highest surface, with its covering of conglomerates, silcretes and gravels, is preserved on both sides of the poort at a height of approximately 1,200 feet to the north and approximately 1,100 feet to the south. Additional evidence is afforded by the cross-section and the trend of the river within the poort. The walls of the gorge slope gently to a point almost level with the higher bench and gravels to the west and east of the poort at a height of 1,200 feet. Below this point the slope becomes steeper, to a point corresponding to the height of the lower (approximately 750 feet) semi-consolidated gravels. Below this it becomes almost vertical for about 520 feet. The course of the river within the mountain is relatively straight, with only very minor and youthful adaptation to structure. This seems to suggest that the river, before its inheritance, had developed a relatively straight course through the mountain and that the surfaces to the north and south of the mountain are of similar age. With uplift the slopes receded sufficiently to reflect the lower stance, before a continuation of the downcutting.

What of the earlier erosion base of the river between the Langeberg and Swartberg? During the late-Cretaceous times deposits collected in the area to the south-east and north of the poort. The Mossel Bay patch (to the south-east of the poort) is bounded by a normal fault to the north, which has tilted the formations to angles of about 20° (see Potgieter, 1950). The lithology of this patch of Cretaceous deposits has not been described in detail, but it is interesting to note that in his description of the area, Haughton (1935) shows that some of the components of the deposits are derived from a northern provenance. Thus, where they lie in contact with Pre-Cambrian rocks, boulders of granite, gneiss and slate are found (p. 16-17). However, "on Hemelrood (A 2) the conglomerates near the village of Herbertsdale dip at angles up to 20° to the north. On the west bank of the river the deposits are silicious, with pebbles and boulders of quartzite set in a ferruginous sandy matrix that imparts a red colour to the cliffs; but on the road leading north from the village, the conglomerates consist almost entirely of muffin-shaped pebbles of Bokkeveld shale in a greenish argillaceous matrix". If the provenance of a part of these deposits also lay to the north, it would seem that a certain significance could be attached to the above description, as the area immediately to the north is composed almost entirely of arenaceous material which is not reflected to the extent one would expect. If, however, the Gouritz Poort was in existence at this time, or the provenance was entirely from

the south and west, then the presence of the shale pebbles and the argillaceous matrix could more easily be accounted for. However, with the lack of detail and the possibility that the material was derived from other parts of the basin rim, the writer feels that no positive evidence can be gained from the above, but considers that the lithology does not contradict the postulate that the poort existed during this period.

Evidence from the Cretaceous deposits north of the Langeberg is confined to the Oudtshoorn basin and to the area around the town of Touws River. This latter area reflects a surface which must have existed during that time. The former area presumably had an eastward slope which was basined by compressive cross-folding in the east disintegrating the drainage (see du Preez, 1944). No evidence is led by du Preez to postulate a similar compressive cross-folding in the west and thus the axis shown on his map, running north-south through Zandberg and Rooiberg, must rather represent a monoclinal axis, to the east of which faulting reached a maximum, deepening the already existing trough. Thus no dismemberment of the river systems is indicated in this region by this "cross-folding", and the river which occupied the valley to the west may have continued to flow into the Oudtshoorn basin. There appears to be no evidence of this either from the deposits in the basin or relic surfaces, and thus the river, in all probability, flowed through the Langeberg.

The Gouritz Poort is definitely not a superimposed river-poort, as Taljaard has pointed out (1948, p. 13). His reasoning is that the only possible surfaces from which it could be superimposed, after the folding of the Cape and Karroo Systems, are the Cretaceous and the "Jura" surfaces. The former, we know, never covered the area, as the deposits were laid down in disconnected "sag-and-fault basins", while the Triassic lavas preserved in the Suurberg area show that erosion had already begun to etch out valleys before the "Jura" surface was developed. The origin of the poort can thus be best postulated as being analogous to the many others in the area, i.e. as a subsequent poort of late-Cretaceous age.

Well established drainage lines must have existed in a region of mature valley sculpture during the formation of the higher surface to the north of Gouritz Poort, as from both the Slang River area to the east (1,800 feet) and from Sandkraal (1,500 feet), Brand River (1,950 feet) and Lemoenshoek (2,200 feet) in the west, the surface underlying the deposits slopes regionally toward the poort. The fact that "back-slopes" are encountered, and the existence of ridges dividing the groups of remnants, raise the problem of the nature of the surface near the mountain. Two possible explanations readily suggest themselves; that the primary surface was flat and that the "back-slopes" were caused by later movement, or that the original surface was gently undulating while being reduced to a number of local stream levels. The fact that the present divides parallel those existing previously, is not significant; but the fact that ridges did exist near the mountains at least (e.g. Lemoenshoek-Brand River), on either side of which surfaces and deposits formed, suggests that low "watersheds" are the more likely.

Presumably the higher surface was not so well developed in the Barrydale-Montagu area as in other parts, but that some development had taken place is shown by shoulders on the mountain slopes rising to over 3,000 feet in the Montagu area. Remnants to the south of Anysberg and the country around Groot Vlakte beacon show that the higher surface was formed at about 2,500 feet and the lower surface a few hundred feet below this. The relatively flat country around Touws River town which existed earlier must also be considered as having been lowered to its present height (2,600 to 2,700 feet) by the rivers during the formation of the higher surface. Little subsequent incision has taken place from this surface.

Before discussing the age of the Buffels River Poort, one statement must be stressed: it is only in a region of mature valley sculpture, with suitable gradients, that the formation and preservation of stream-cut benches, and of river-lain gravels on the benches, can take place. Neither in youth nor in old age can such a feature be formed and preserved.

The fact that neither benches nor gravels, corresponding to the higher surface, could be found in the region between the Klein Swartberg and the Witteberg, and that today the region shows many features of youth, strongly suggests that during the earlier period it was in a youthful stage of development. Remnants of the lower surface are found near the Buffels River at a height of about 2,000 feet, which can be correlated with the larger terraces to the south of Laingsburg on the northern flank of the Witteberg (2,500 feet). As has been stated earlier, this intermontane area is more suitable for the preservation of any remnants had they been formed, than the area to the south of the Klein Swartberg where they are preserved.

It seems from the above, therefore, that breaching of the mountain took place between the two periods during which surfaces were formed. Further confirmation of this may be gained from an examination of the cross-section of the poort. The lower 300 to 400 feet of the walls are almost vertical and above this point, which corresponds to that of the lower surface to the south of the poort, the slope of the walls decreases sharply. A further decrease in slope at a height comparable with that of the higher surface, such as is displayed by the section of Gouritz Poort, is not apparent. With regard to the course of the river within the mountain, which resembles that of a meander flowing over flat ground, it must be considered as an anomaly. A result of accepting the date above is that the poort must be subsequent. The direction of the tributaries, which are presumably strongly controlled by the structure of the rocks, is parallel to those of the main stream. The turns in the centre of the mountain display slip-off and under-cut slopes at heights near those of the remnants of the lower surface to the south. This phenomenon is significantly situated near the anticlinal axial plane. An explanation of the apparent anomaly may thus lie in its relation to the position of the axial plane, the joint pattern, the resistance of the rocks and the nature of the entrenchment. The latter was presumably slow enough to permit the formation of slip-off slopes in the beginning and must have been followed by rapid incision. This cannot be considered a meander as the structure determines the form. This last incision has produced a number of discordant valleys.

There is no evidence of drainage during the earlier period in the area to the north of the Witteberg (3,000 feet), if this is represented by the highest surface in the area. It is interesting to note that the lowest present-day watershed to the east is situated between 2,500 feet and 2,750 feet (topographical map), just north of the Doekberg: the divide of the Hartebeestspruit and Jakhals River.

The Ladismith region up to and including Amalienstein (3,000 feet) and Vaartwel (3,100 feet) presumably drained to the west, if the higher surface is undisturbed. Seven Week's Poort must have attained considerable development by this time if the evidence of the slopes of the walls at the exit of the poort is to be believed. They are as low as 20° to 30°, below which they steepen abruptly. However, stereoscopic investigation of the upper reaches suggests that the present-day lines were inherited from the upper surface with little modification and there are no gravel remnants. The lower surface is not developed to any extent in this area, but it may be represented by a few shoulders and the tops of the ridges, which have a slight gravel covering. If, as is suggested above, the drainage was originally to the west, capture of the rivers in this area by the Huis River must since have taken place.

The terraces and benches around the north and west of Rooiberg show that the fall was to the west in the north (2,500 feet to 2,000 feet), but the group of terraces on the southern flank must be considered to be the remnants of the lower surface because of their heights (1,404 feet at Vanwyksdorp and 1,100 feet in the east near Jagtberg Poort) and their unconsolidated nature. The same applies to the bench along the left bank of the Gamka River in Jagtberg Poort (approximately 1,200 feet). Of interest is the fact that this bench is only developed along the Gamka River and that the portion occupied by the present Olifants River, above their confluence, shows no equivalent development, save possibly a smooth structural bench 200 feet above that on the other "fork" which dips in an opposite direction to the present river. The regional slope of the shoulders on the southern bank opposite the bench mentioned above, also appears to be to the east. It seems probable, therefore, that the course occupied by the Olifants River between Jagtberg and Gamka Hill, above its confluence with the Gamka River, is of more recent origin, i.e. post-lower surface. The course of the Gamka River between the Rooiberg and Jagtberg-Gamka Hill must be of pre-higher surface age, as is shown by the fact that the surface is well developed both to north and south of the poort.

From the height of the remnant to the north-west of the poort this higher surface must have crossed the range at a height of about 1,600 to 1,700 feet, and the valleyward dips that can be seen, at Kerkrand for example, suggest a more northerly course of the Olifants River. The slope during the beginning of the Cretaceous deposition in this area was eastward and after the cross-folding, centroclinal, thus proving that the Jagtberg (Gamka portion) breaching must have occurred between the late-Cretaceous and the higher surface periods. This may be associated with the cessation of the deposition in the area.

A similar date can be ascribed to Gamka Poort through the Swartberg-Gamkasberg.

All the remaining poorts in the area, namely Prins Poort, Kogmanskloof, Tradouw and Garcia's Pass, suggest that their breachings occurred later than the lower surface, with the possible exception of Kogmanskloof which may have been slightly earlier.

The breaching of Aasvoëlberg, the ridge to the south of the Albertinia-Mossel Bay road, may be considered here, even though it falls outside of the area investigated. The fact that the remnant terraces, covered with consolidated deposits, dip towards the present river must be interpreted to mean that the river afforded a local base level during the formation of the surface, and that it drained through the ridge during this period. The deposits of Enon to both north and south prove that during that time a disintegrated drainage also existed in this area, and thus the breaching must best be regarded as similar in age and origin to that of the Gamka Poort, i.e. a fault scarp stream of post-late-Cretaceous, pre-Eocene age.

Summing up, the writer finds that all the poorts in the Western Little Karroo are presumably of subsequent origin, that is, they developed by headward extension, and that their dates of breaching can be tabulated as follows: (Dates of surfaces, see later).

One result of this dating shows that the deduction of Rogers, that the Buffels River once flowed over Garcia's Pass, falls away because the earliest possible date for such a course must have been during the Cretaceous age, as is shown by the higher surface. At that time the Buffels River did not exist.

Quaternary	Recent		
	Pleistocene		Prins, Garcia's, Tradouws, Jagtberg (Olifants, portion), etc.
Tertiary	Pliocene	Lower Surface	Kogmanskloof?
	Miocene		Buffels
	Oligocene	Higher Surface	
	Eocene		Aasvoëlberg, Gamka and Jagtberg (Gamka, portion).
Cretaceous	Late	Enon deposits	Gouritz.

XI THE SURFACES AND OVERLYING DEPOSITS

Higher Surface usually with consolidated deposits.

The remnants of this surface in the western and north-western parts of the area (Grootvlakte and Touws River), which have no resistant protective covering, have been reworked and lowered subsequent to formation. In the central and eastern parts of the area, however, remnants flank the mountains and the original surface has been preserved under consolidated deposits in a number of places.

This surface, in a valleyward direction, is concave, as has been shown by the measurements in the areas near Lemoenshoek, Long Mesa, Sandkraal, Waaikraal, and north of the Touwsberg. They vary between 300 feet/mile near the mountain, to 50 feet/mile towards the valley. There is little doubt, however, that towards the valley the surface had an even lower gradient comparable with the longitudinal inclination (16 to 20 feet/mile, Lemoenshoek-Gourtiz Poort and 10 feet/mile Oudtshoorn basin).

Near the mountain, tongues of the relatively flat surface extended to the mountain front, divided by watersheds. How far these divides extended valleywards above the average height of the surface cannot be stated with certainty, but they may, as in the case of the Lemoenshoek-Brand River divide, have stretched a matter of miles from the mountain range. The surface towards the valley was thus undulatory in transverse profile, to at least this distance.

The deposits which rest on the remnants of this surface, with the exception of the talus which makes an angle with the surface, have a stratification which is roughly parallel to the surface. The nature of the deposits immediately overlying the surface shows that they are foreign to the underlying rocks and thus must have been transported. The rounding of the larger particles (boulders, pebbles, etc.), and the occasional percussion mark, suggest a fluviatile origin. The thickness of the deposits, where overlain by talus but not including it, varies from 25 to 110 feet, averaging about 55 feet throughout the area. The upward gradation from small, rounded pebbles to larger, less well rounded boulders shows an increase in competency. Above this the sand, with its pebble bands and lenses, suggests a decrease and fluctuation in competency, probably brought about by climatic changes.

All these factors indicate that the surface originated under fluviatile conditions and was reduced to the level of the master stream and the tributaries by processes including lateral planation, weathering, rain wash, rill wash and sheet flood. The deposits represent those of aggrading rivers in a mature stage of development with a pattern

similar to those of today and in an area which experienced climatic changes. The talus is regarded as of later date—presumably the result of incision of the streams into and below these deposits.

Consolidation of the deposits by silicious and ferruginous cements occurred under conditions of a fluctuating water table in the subsoil, as has been shown by Frankel and Kent (1937). In a valley containing deposits, the area adjacent to the river will experience little or no change of the ground water table, and thus little cementation. Immediately next to the mountain similar conditions will result, as the movement of water is due to seepage and evaporation and not a fluctuation of the water table. The maximum fluctuation, and thus maximum cementation, will take place near the mountain although a short distance from it. The shape and position of the remnants bear this out: even where they overlie rocks of similar resistance, they are seldom contiguous with the mountain, and show a maximum width some distance from it. Had the cementation been uniform, the opposite effect would have resulted: the maximum width of the remnants would be adjacent to, and contiguous with, the mountain. Another result would have been that the deposits near, and possibly within, the poorts would have been cemented.

The climatic phases postulated above are not out of keeping with the views of other authors (see for example, du Toit, 1939, p. 507). Frankel and Kent (1937) find that in the Grahamstown area "all the original deposits, whether the peneplain is of fluviatile or marine origin, seem to have been removed by unknown agencies, probably aeolian, and the peneplain subsequently weathered for a long period under subaerial conditions, resulting in the formation of incoherent residual deposits" (p. 38). The cementation which then took place in the subsoil, was interrupted on several occasions: "These alternating periods of silification and non-silification are probably ascribed to climatic variation—it has been shown that a seasonal fluctuating water table is essential for the formation of silcretes" (p. 39).

The Lower Surface.

Although this surface was not studied in detail, from the few remnants visited some conclusions can be drawn. The planation of this surface was not as well developed as that of the higher, and the thickness of the unconsolidated gravel mantle seldom exceeds 25 feet. The constituents of the gravel are, as in the case of the higher surface, mainly transported material which have a fluviatile origin, as is shown by the chattermarks. Davis (1906, i, p. 397) wrote of the remnants to the north of the Witteberg as follows: "The reduction of the surface to a plain must have been largely aided by the general wasting down of the minor ridges as well as lateral swinging of the streams. The distribution of the coarse cobbles over the plain seems to have been the work of sheet-floods, such as become peculiarly effective in the later stages of a cycle of erosion, particularly in a region where occasional heavy downpours of rain occur". The writer suggests that the above mode of formation of the surface and the gravels applies to the other parts of the area equally well.

XII AGES OF THE SURFACES

No evidence for accurately fixing dates to either of the surfaces was found in the area under discussion. As they cut across the tilted Enon deposits in the Oudtshoorn basin, they must be at least of post-late-Cretaceous age.

Dates have, however, been ascribed to the higher surface to the south of the Langeberg by various authors, who usually relate it to the Bredasdorp-Alexandria

beds, found nearer the coast. Some suggest they were contemporaneous (e.g. du Toit and Haughton) while others consider that the terrestrial deposits were later than (Potgieter), or younger than (Wybergh) the marine. In passing it should be noted that the age of the Bredasdorp-Alexandria beds, on the European time scale, cannot be fixed with any certainty, being found to be similar to the Eocene (Haughton and Chapman) and the Mio-Pliocene (Newton). (See du Toit, 1939, p. 401).

King (1951) has attempted to correlate cycles of erosion related to a base level. This has resulted in surfaces separated by 5,000 feet belonging to the same "cycle" p. 191 and p. 256). Also a surface of erosion overlying a surface of erosion and deposition, is considered as belonging to the same "cycle" (p. 248). These surfaces may have been formed in different geological Periods. With regard to the higher surface in this area he tentatively suggests that it falls in the "African" cycle (p. 311). This was "initiated in the earliest Cretaceous" (map at end). However, this surface "appears between 2,000 feet and 3,000 feet near the sea, and 3,000 to 4,000 feet in the interior" (p. 251) and in the south of this area it lies at about 1,000 feet and, adjacent to Aasvoëlberg at less than 500 feet.

In an attempt to date the surfaces, the writer assumes that the coastal limestone deposits south of Aasvoëlberg are of Mio-Pliocene age, and that the 20-foot marine benches and terrestrial terraces are of Pleistocene age.

Wybergh (1919, p. 51) states: "It would appear, therefore, from the wide distribution of the pebbles and their peculiar mode of occurrence that they were probably already distributed over the surface of the area upon which the Bredasdorp beds were laid down, but were 'worked over' again and incorporated with the latter, receiving at the same time the marine shells which they contain. The pebbles themselves probably date back to a much earlier period".

Basing the argument on the assumption above and Wybergh's conclusions, it would appear that the surface underlying the higher, consolidated deposits extends under the Bredasdorp beds in this area. This must date it as pre-Miocene and, from its attitude to the Enon beds, post-late-Cretaceous, i.e. Eo-Oligocene. The lower surface was thus, in all probability, the result of the incursion of the sea, and thus of Mio-Pliocene age.

XIII NATURE OF THE UPLIFT

During the Tertiary period many parts of the sub-continent experienced differential uplift. Du Toit has indicated the regions of depression and the axes of uplift (1933). Those of the coastal regions are possibly associated with continental sliding and the origin of submarine canyons (1940).

In the Baviaanskloof area, de Villiers (1942, p. 144) has shown the existence of an east-west trending axis of upwarping of post-Eocene age, situated in the Couga Mountains. He suggests that a similar type of uplift occurred in the Kamanassie Valley, but does not suggest any locality.

To the south of the Langeberg, Taljaard (1949, p. 78) has shown the existence of another axis which he has followed from a position south of Swellendam to east of Riversdale. The warping was so intense that a back-slope has been produced in some places.

Evidence from the area to the north of the Langeberge which also suggests warping is fivefold:—

1. The relatively narrow valleys of the Slang River, the Kamma River, the Hell and those between the Touwsberg and the Swartberg, all have a maximum distri-

bution of remnants of the Eo-Oligocene surface to the south of the centre of the valley. They have been completely removed to the north in some cases. A similar condition is displayed by the Mio-Pliocene surface between the Touwsberg and the Swartberg.

2. In the more open valleys, such as that between the Langeberg and the Rooiberg, that between the Warmwatersberg and the Anys-Touwsberg, and the Oudtshoorn Basin, the rivers have a tendency to flow either to the north (first two) or the south (Olifants River) of the centre of the valley. The same may be said of the Slang River, the Groot River (between the Touwsberg and the Swartberg), and other rivers in the area which are situated to the north of the centre of the valleys.

3. The depth of incision, too, may be an indication of the tilting nature of the uplift. In the area to the north of the Langeberg, for example, the incision from the Eocene surface is between 300 and 500 feet, whereas in the region to the south of the Swartberg it is considerably more, varying from 600 to 900 feet. The area around Jagtberg shows incision of about 800 feet from the higher surface. From the lower surface to the present-day rivers there is a variation in the figures obtained from the different parts of the area. Almost everywhere it is about 300 feet, but in the area near Gouritz Poort, the Rooiberg and Jagtberg Poort it is between 500 and 600 feet. When considering these figures it must be remembered that the present rivers have not attained the low gradient possessed by those of the earlier periods.

4. The differences in elevation between the two surfaces, too, suggest that warping has taken place in certain areas. Between the Touwsberg and the Swartberg, for example, the difference in elevation in the south is about 100 feet, while in the north the difference is over 600 feet. It has already been mentioned that near Long Mesa the "lower" surface appears to be above the "higher".

5. The meanders displayed by some of the rivers of the Gouritz River System around Gouritz Poort may be suggestive of unequal uplift.

These points evince that warping has occurred in this area. The exact position of the hinge lines and the date at which they were formed are still vague, but some idea of these may be gained from an examination of the diagrammatical sections of certain of the valleys (Fig. 8). These are based on points 1 and 2 above; they are not drawn to scale.

The section from Touwsberg to the Swartberg (A-B on the map) appears to be very similar to that of the lower portion (up to the Eo-Oligocene surface) of the Baviaanskloof (I-J on the map. From de Villiers, 1938, Fig. 16). It has been suggested by de Villiers (1938, p. 143) that the Baviaanskloof experienced two successive periods of surface formation, each followed by a northward tilting of the area and incision of the rivers. A similar sequence of events may be postulated here. The variation in the difference in elevation between the higher and lower surfaces (point 4 above) may be explained by the movement of the local erosion base and possibly also by the breaching of Buffels River Poort.

The valleys of the Hell, Slang River and Kamma River (sections C-D) suggest that they too were tilted at some period after the formation of the Eo-Oligocene surface. The lower surface was not studied in these areas.

From the positions of the areas described above it seems reasonable to suggest that if one axis of uplift is postulated, it lies to the south of the Langeberg. This, however, is contradicted by the amount of subsequent incision, for where the greater uplift occurs, i.e. in the southern parts, the greater incision must result, provided the rivers have similar gradients. The remnants against the northern flank of the Langeberg, for example, are some 10 miles nearer the common erosion base than those

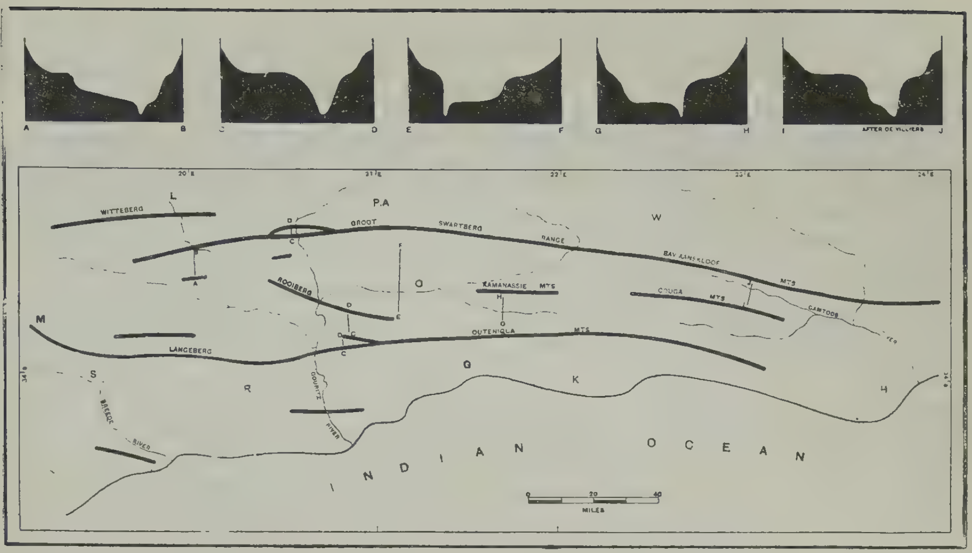

Figure 8

Diagrammatic Valley Sections.

G = George. H = Humansdorp. K = Knysna. L = Laingsburg. M = Montagu. O = Oudtshoorn. P. A = Prins Albert. R = Riversdale. S = Swellendam. W = Willowmore.

against the southern flank of the Swartberg. The total incision from the higher surface, however, is greater in the north than in the south: against the Langeberg it is 870 feet (Gouritz Poort), approximately 400 feet (Long Mesa), against Rooiberg about 800 feet (Jagtberg Poort), to the north of the Touwsberg about 650 feet and to the south of the Swartberg 800 to 900 feet. [The corresponding figures for the lower surface are 512 feet (Gouritz Poort), 400 feet (Long Mesa), 600 feet (Vanwyksdorp), 350 feet (Touwsberg) and 300 feet (Voorbaat)].

It seems likely, therefore, that at least two separate warpings have occurred (post-Eo-Oligocene surface and post-Mio-Pliocene surface). Presumably they did not have the same axis. The amount of uplift along each axis may have been differential.

The key to the problem seems to lie in the Kamanassie Valley. In this area de Villiers (1936) found that the higher surface has a maximum distribution on the northern side and the lower surface on the southern side of the river (see section G-H, Fig. 8). This would suggest that uplift after the formation of the higher surface resulted in a southerly migration of the river, while the movement succeeding the lower surface caused the rivers to migrate in the opposite direction, that is, northwards.

A possible explanation of the movements in the area to the west of the Kamanassie Valley, may be as follows: After the formation of the Eo-Oligocene surface, warping occurred. This warping was an extension of, although not necessarily continuous with, the line of warping which affected the Baviaanskloof and Kamanassie areas. In these areas the rivers were caused to migrate to north and south respectively, removing the deposits in those directions. This hinge-line in all probability caused the

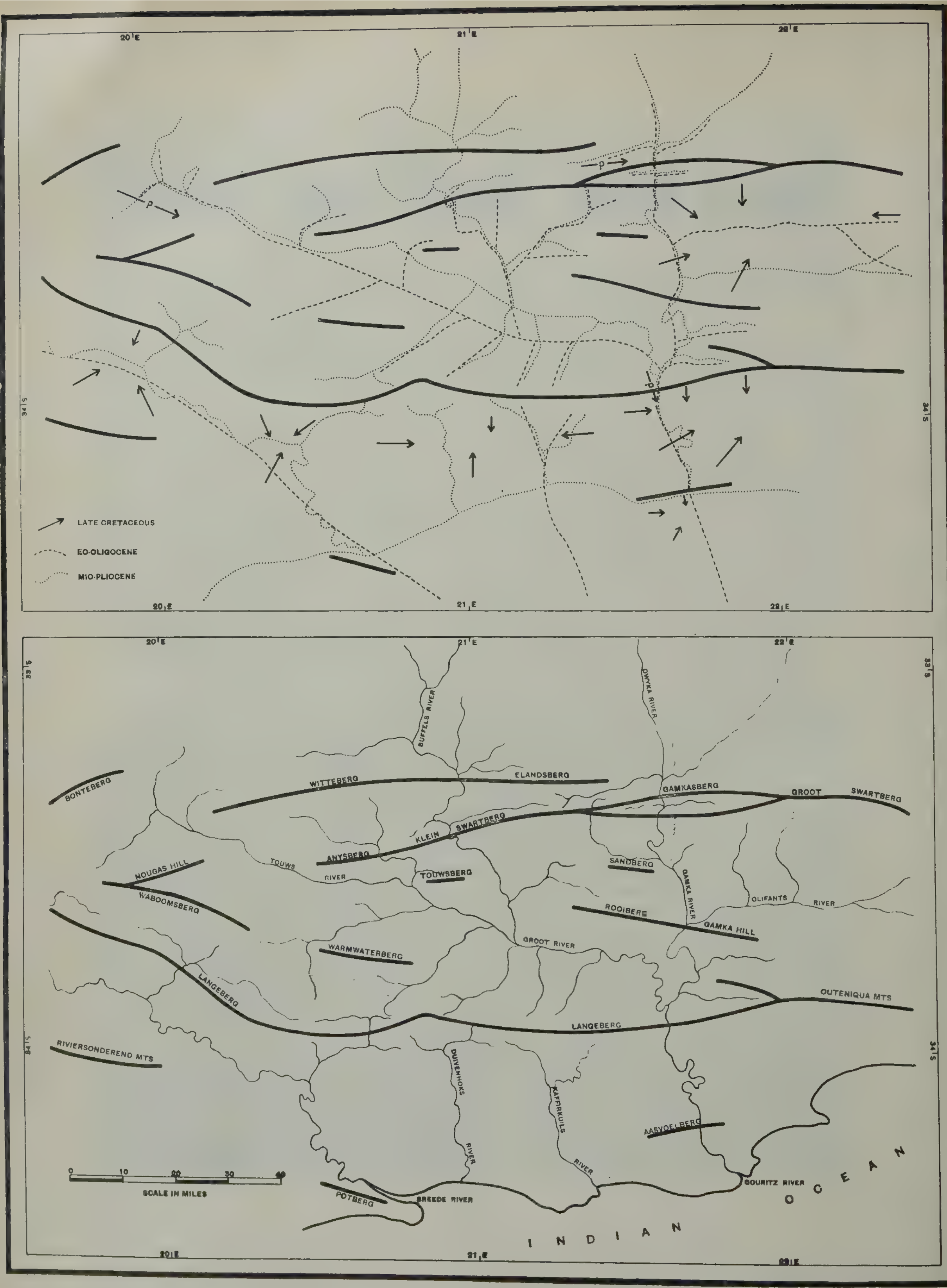

Figure 9

Above: Drainage Development. (Late-Cretaceous to Mio-Pliocene)
Below: Presentday Drainage.

Olifants River to migrate southwards in the Oudtshoorn basin and the Groot River to migrate northwards in the area north of Touwsberg, again removing the gravels in those directions. The area to the south remained relatively unaffected by these movements; they were in all probability elevated reasonably evenly. The second (post-Mio-Pliocene) tilting presumably then caused the Groot-Touws and Kamanassie Rivers to migrate to the north. The Slang and Kamma Rivers which had remained virtually unaffected as regards position by the first movement, then moved northward. This second axis must lie to the south of the Langeberg and may have coincided with that traced to the south of the towns of Swellendam and Riversdale. Another result of this later warping was the formation of the meanders displayed by the rivers for some 30 miles north of the warping axis. They were subsequently entrenched to approximately their present position. The Gouritz River shows abnormally large entrenched meanders north of an extrapolation of the warp axis, and to the south a remarkably straight course.

Another point which may be noted is that the rivers to the north of the Swartberg have been relatively unaffected by the tilting in the Buffels River Basin, while a number of the subsequent tributaries of the Gamka Basin have asymmetrical valleys.

The Olifants River has remained near the southern range. This may be due to the nature and attitude of the Enon deposits over which it flows.

XIV THE DRAINAGE EVOLUTION OF THE GOURITZ RIVER SYSTEM

(*Based on du Toit* (1939), *du Preez* (1944), *de Villiers* (1942), *Taljaard* (1949) *and the present paper.*)

The birth of the major drainage lines in the Southern Cape Fold Belt occurred at the beginning of the Mesozoic Era. Very little is known of the direction of the drainage before this time, but it was, in all probability, northwards from this area. Even though the evidence from the time of inauguration is very sketchy, there seems to be little doubt that an east-west trending drainage existed prior to the Triassic Period. This is suggested by the attitude of the "Suurberg lavas" in the Uitenhage area, which rest in an eroded, structural valley. It is very doubtful whether the surface attributed to the late-Jurassic Period has any relics in this area, and thus it may be presumed that it did not obliterate the strike ridges of the folded formations.

Before the late-Cretaceous Period these valleys had been excavated, so that in the intermontane valleys, such as the Oudtshoorn area, a high relief must have existed. At about this time a regional, differential, small scale submergence of the continent occurred and the sea-board moved south-westwards. In the interior this movement consisted of a basining of the valleys. This was accomplished by sagging and cross-folding, followed by normal fauiting. In these centroclinal, structural and tectonic basins deposits began to accumulate. Certain parts of the surface which existed at this time were thus buried, while the remainder were eroded. Today reflections of this latter part of the surface can be seen around Touws River town, on the Klein Swartberg, and on the northern flank of the Swartberg to the west of Gamka Poort. The climate suggested by the nature of the deposits must have been similar to that of the Oudtshoorn area, namely semi-arid. The drainage at this time was thus disintegrated, flowing into various basins, and there is a possibility that Gouritz Poort came into

being at about this time. (See Fig. 9 for drainage lines). (Du Toit has suggested that the separation of originally continuous deposits occurred. The above sequence of events, as suggested by du Preez, seems to be more likely.) The sea-shore at this time must have been somewhere to the south and east in the present Indian Ocean.

Uplift, presumably of a differential nature, then followed, elevating the marine deposits, and causing a change in the position of the sea-board. At this time, possibly associated with the movement, basaltic magma was injected into the crust (melilite basalt and kimberlite). The increase in the erosive power that the rivers experienced as a result of the uplift, may have been the reason for the breachings of Aasvoëlberg, Rooiberg-Gamka Hills and the Swartberg which occurred at this time. The major rivers and tributaries reduced their gradients considerably, until a mature stage of valley development was attained in the area. A possible scheme of the drainage lines can be seen on the accompanying map. On the flood-plains of these streams, deposits accumulated to considerable depths. These deposits reflect climatic conditions very similar to those of the earlier period of deposition, namely a predominantly semi-arid climate with phase variations. It is possible that at least two of the later phases were such that a seasonally fluctuating ground-water table was produced. Under these conditions colloidal silica and iron oxides were leached from the underlying formations and cemented the overlying deposits.

Uplift along an axis then occurred which in this area is shown by the nature and relationship of the remnants of the surfaces in the area between Touwsberg and the Swartberg. The river in this area, the Groot, most probably succeeded in breaching the mountains to the north and formed Buffels River and Leeukloof Poorts at this time.

At the same time the sea advanced as far as Aasvoëlberg. It reworked the river deposits and then covered them with a considerable thickness of shelly limestone deposits. These are suggested as being of Mio-Pliocene age. (The fact that the underlying gravels show no signs of cementation and that no pebbles of silcrete have been found in the flood deposits in the marine beds, suggests that the cementation had not taken place by this time.)

In the hinterland the Eo-Oligocene Epoch surface became dissected and scree began to accumulate on many parts of the surface which was not kept cleared by the streams. To this new base-level the rivers cut a new surface which carried a thin mantle of gravels. The drainage lines at this time can be seen on the accompanying map.

Once again uplift occurred, again of a warping nature. To the north of the axis the rivers were dammed up and forced to meander, or if they already possessed meanders, these were enlarged. To the south, as a result of the increased gradient, they tended to straighten their courses. As the strand line had moved further south by this time, the rivers were extended in consequent manner over this newly-formed surface. Doubtless as a result of the uplift and subsequent increase in erosive power, certain rivers breached the Langeberg (Tradouws and Garcia's Passes.) The breachings at Kogmanskloof and Prins Poort may have occurred at about this period, but the occasional gravel in their drainage basins to the north of the mountains suggests a slightly earlier date. The drainage lines thus became very similar to those seen today (see Fig. 9).

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THE MINERALOGY AND GENESIS OF THE LEAD - ZINC - VANADIUM DEPOSIT OF ABENAB WEST IN THE OTAVI MOUNTAINS, SOUTH WEST AFRICA

By

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(Submitted March, 1953)

ABSTRACT

The economic ore minerals in the oxidized Abenab West ore body are descloizite, vanadinite, cerussite and galena, all of which are distributed in an unconsolidated ferruginous clay along a disturbed zone in dolomite and limestone. A shallow reef of drusy, secondary willemite is situated at the eastern end of the mineralised clay, constituting an economic source of zinc ore. In the scattered remnants of primary ore small amounts of sphalerite, pyrite, bournonite, tennantite and chalcopyrite were identified, while the minor secondary ore minerals are anglesite, covellite, greenockite, mimetite, smithsonite, malachite and wulfenite. The gangue minerals are "limonite", goethite, montmorillonite, kaolinite, calcite, quartz and a number of accessories. A few minor constituents could not be identified, while a major portion of the alumina and zinc in analyses of the ore is not accounted for by the known minerals.

Mineralogically, the descloizite, vanadinite and willemite present several interesting features that are discussed at some length. The mode of occurrence of the principal ore mineral, descloizite is given in detail.

From a systematic mineralogic study of mine samples, three concentric vanadinite-descloizite zones could be distinguished in the ore body, with vanadinite predominating towards the centre and descloizite towards the periphery. These are superimposed on a twofold core of galena-cerussite, characterised by an increased lead content and a concomitant decrease of vanadium. Willemite has a well-defined marginal relationship.

The ore body is considered to be the oxidized and extensively contaminated hood zone of a sulphide vein that had a low copper content but must have carried appreciable pyrite. The sulphides had probably crystallised under mesothermal conditions and were subjected to subsequent deformation. The oxidation of the ore body took place mainly during an arid climatic period that is held responsible for the characteristic suite of secondary minerals. The vanadium mineralisation was one of the last events in the geologic history of the ore body and is separated from the sulphide mineralisation by a long period of time. Two generations of vanadinite and mimetite are distinguished in the Abenab West and surrounding deposits, the earlier vanadinite having undergone solution and extensive replacement by descloizite. Pseudomorphs and casts of several other minerals indicate that the oxidation of the ore body was characterised by a series of transformations and replacements due to changing supergene conditions. Slumping and deposition of bedded clay and sand played an important part in the formation of the ore body, both before and after the deposition of the vanadates. The alterations effected during the present-day humid oxidation period are insignificant compared with those effected beforehand.

There is little doubt that the vanadates were deposited by supergene weathering solutions. New evidence of the conditions of deposition of the vanadates of the Otavi Mountain Land, including spectrographic data and the results of decrepitation tests, appears to support this conclusion. The most likely source of the vanadium is thought to be the shale horizons in the Otavi System, the leaching of vanadium having taken place during the arid oxidation period most probably from shale that was totally decomposed and afterwards removed by erosion.

This paper was awarded the Junior Scott Memorial Medal for the year 1954.

CONTENTS

I INTRODUCTION

The Abenab West mine is at present the major producer of vanadium ore in the Otavi Mountain region of the Grootfontein district, South West Africa. It is situated about 300 yards to the south-west of the famous Abenab mine (now closed) which was probably the largest known occurrence of vanadate ore.

Though prospecting operations at Abenab West commenced as long ago as 1924, the comparatively low vanadium content of the ore did not permit exploitation before 1939. Mining proceeded intermittently until May 1947 when new reduction installations rendered the systematic concentration of the ore possible. Up to the 25th October, 1952, 27,926 tons gravity concentrates averaging about 58% Pb and 5% V_2O_5 and 6,527 tons flotation concentrates averaging 51% Pb and 13% V_2O_5 have been produced.

Geologically and mineralogically the ore body presents several interesting features. Although the proximity of the Abenab and Abenab West deposits suggests a genetic relationship, the difference in structure, form and mineralisation between the two is striking. The Abenab ore body, on the one hand, consisted of a pipe-like mass of brecciated country rock, cemented by compact reddish "clay" and coarsely crystalline calcite, descloizite and vanadinite, which occasionally formed beautiful drusy cavity linings. The Abenab West deposit, on the other hand, is a reef of unconsolidated ferruginous clay in which the presence of minutely disseminated vanadinite and descloizite could hardly be suspected at first sight. This clay-like ore body, situated along a minor fault zone in dolomite, also carries lumps of partly altered galena and is thought to represent the oxidized and contaminated hood zone of a sulphide body which is presumably continued in depth. The colour, size, habit and associations of the vanadium minerals appeared to differ from those recorded elsewhere, while the apparently zonal distribution of the ores also aroused interest. In addition, it was discovered that a particularly fine occurrence of secondary willemite is developed at the outcrop. There are several points of resemblance between the Abenab West ore body and that of Broken Hill, Northern Rhodesia, which probably belongs to the same mineralisation epoch.

The present research was undertaken at the request and with the co-operation of the South West Africa Company, Limited. The major object of the investigation was to establish the mineralogical composition of the ore body in all parts of the mine, a knowledge of which was considered indispensable for the proper recovery of the valuable constituents. In addition it was hoped that useful information might be obtained on the mineralogy of the descloizite-mottramite series, and that new light might be shed on the problem of the origin of the vanadium in vanadate ore deposits.

The work was conducted in the Department of Geology and Mineralogy of the University of Stellenbosch on a large number of specimens and representative samples from the Abenab West mine, collected by the author during his stay at Abenab from December, 1950 to January, 1951. Attention was also given to the minor, but closely related vanadate and sulphide occurrences in the mining area. All available specimens of vanadium ore collected earlier by Dr. C. M. Schwellnus, Dr. J. W. Brandt and others from various localities in the Otavi region, and especially from the old Abenab mine, were included for comparative study.

II PREVIOUS LITERATURE

The occurrence of vanadium minerals in the Otavi Mountain Land has been known since 1908 when Maucher first mentioned the presence of mottramite in the oxidized zone of the Tsumeb copper mine. (Maucher, 1908.) Subsequently, numerous small deposits were discovered in sand-filled solution hollows in the Otavi Dolomite, as well as a few larger ones associated with ores of copper, lead and zinc like those at Abenab West and Berg Aukas. The mode of occurrence and geological features of these deposits have been discussed by A. Stahl, H. Schneiderhöhn, A. W. Clark, and more recently and comprehensively by C. M. Schwellnus who was a senior member of the field party responsible for the geological survey of the Otavi area in 1940. A Geological Survey Memoir dealing with "The Geology and Vanadium Deposits of the Otavi Mountain Land" has not yet been published.

Discovered in 1921, the large Abenab deposit was first described in some detail by A. W. Clark (1931) who also mentioned the occurrence at Abenab West. The latter, forming the main subject of this investigation, has been described by C. M. Schwellnus (Willemse, Schwellnus *et al.*, 1944; Schwellnus, 1945).

The various geologists who studied the vanadium deposits reached widely different conclusions as to the genesis of the ore. The observations of these investigators bearing on the genetic problem are summarised in a later chapter. P. A. Wagner (1922) and A. Stahl (1926) regarded the vanadates as due to the secondary enrichment, by ground water, of small amounts of vanadium in originally overlying deposits of lead, zinc and copper sulphides, presumably richer in vanadium than the portions remaining today. H. Schneiderhöhn (1929) expressed the opinion that the vanadium is of biogenic origin and was transported by surface waters from old land surfaces to oxidizing sulphide deposits. A. W. Clark and C. M. Schwellnus concluded that the vanadium was leached from the sediments of the Otavi System by meteoric waters and deposited as metal vanadates at the sub-outcrops or in the vicinity of oxidizing lead-copper-zinc ore bodies. A similar conclusion on the origin of the vanadium ores of Broken Hill, Northern Rhodesia, was reached by A. C. Skerl (1934). H. Korn and H. Martin, in a private report to the S.W.A. Company (1937) quoted by Schwellnus (1945), suggested the hydrothermal deposition of the vanadates, and this opinion was shared by some members of the 1940 field party. The hydrothermal theory has recently gained in favour with the geological staff of the South West Africa Company

The mineralogy of the vanadates from the Otavi Mountain Land first attracted the attention of O. Pufahl (1920), who briefly described the external characteristics and chemical reactions of mottramite from Tsumeb; he gave three new analyses and described his analytical procedure. P. A. Wagner (1922) gave a more extensive account of descloizite from Abenab and Olifantsfontein West. He distinguished four varieties of descloizite at the former locality, differing in habit, colour, pleochroism and apparently in copper content. He also noted the zoning and pleochroic variations in single crystals and suggested that we are dealing with at least three isomorphous substances instead of a homogeneous mineral. A. Diefenbach (1930) undertook a mineralogical and chemical investigation of 32 specimens of descloizite and mottramite collected by Stahl and Schneiderhöhn prior to 1924 at various localities in the Otavi Mountain Land (excluding the Abenab area). Ten new analyses were given and discussed, the results of a thorough morphologic crystallographic study were presented and attempts to determine some of the optic constants were described. Diefenbach studied the microscopic features of his specimens with the aid of polished and thin sections, recognised the presence of growth zones, but did not investigate the internal

structure of descloizite in detail. E. Dittler and H. Hüber (1931) analysed specimens of descloizite from Abenab and cuprian descloizite from Olifantsfontein and described their crystallography for purposes of comparison with mottramite from Bolivia.

The mineralogy of descloizite and vanadinite from Broken Hill, Northern Rhodesia, has been briefly described by L. J. Spencer as far back as 1908 (Spencer, 1908) and is similar to that of the South West African occurrences.

Several unsolved problems in connection with the mineralogy of the descloizite family were clarified in a paper by F. A. Bannister (1933). He proved the structural identity of descloizite, cuprodescloizite, mottramite, psittacinite, chileïte, eusynchite and dechenite by X-ray study of material from the type localities. Thus an isomorphous relationship between the zinc-rich and copper-rich end-members of a series was established, as previously suggested most clearly by Dittler and Hüber (1931). Bannister determined the space group and unit cell dimensions of descloizite and gave the following structural formula to the series:

$$4[Pb(Zn, Cu)\ (OH)\ (V, As, P)O_4]$$

He recommended that the name descloizite should be retained for the zinc-rich varieties containing up to two atoms of copper per unit cell (i.e. a content of less than 10% CuO) and mottramite for copper-rich varieties with more than two atoms of copper per unit cell. In accordance with the terminology in Dana's System of Mineralogy (Palache, Berman and Frondel, 1951) the adjectives *cuprian* and *zincian* are used in this paper for further qualification.

Among the specimens analysed and described by Bannister were three crystalline and botryoidal mottramites from various localities in the Otavi Mountain Land, and one black pyramidal descloizite from Abenab. Approximate optical determinations on these minerals were made.

Since 1933 no attempt seems to have been made to re-study these minerals of which beautiful museum specimens are found in every large mineral collection in South Africa. Some of them are wrongly labelled as psittacinite (Schwellnus, 1945, p. 57) which is a discarded name for mottramite. So-called "volborthite" collected by Schwellnus proved, on chemical investigation, to be mottramite also.

III GEOLOGICAL SETTING

The author is indebted to Dr. J. W. Brandt, chief geologist of the S.W.A. Company, for the following note on the geology of the Otavi Mountain district:

"The Otavi Mountain Land is a group of rugged karst-land ranges consisting mainly of folded arenaceous, dolomitic and argillaceous rocks collectively known as the Otavi Dolomite. It contains a great number of lithological units and is of the order of a system. In tabular form the Otavi System is sub-divided into the following descending stratigraphical sequence:*

Economic quantities of lead, zinc, copper and vanadium ores are confined to the Otavi System. The stratigraphical position of the more important deposits are shown in the table."

The Otavi System rests unconformably on a basement of archean schists and granite which is extensively buried below a cover of recent sand and surface limestone. The red eolian sand encroaches on the Otavi Mountains from the east and floors

* Established by co-operation of the geological staffs of the South West Africa Co. Ltd. and the Tsumeb Corporation Ltd.

TABLE 1: SUB-DIVISIONS OF THE OTAVI SYSTEM

Mulden Series			Quartzite and arkose Shale and Slate	
Dolomite Series	Tsumeb Stage	Upper	Thin-bedded light dolomite with oölitic chert.	Tsumeb mine
		Middle	Thin-bedded dark dolomite and dark limestone.	
		Lower	Thick-bedded dolomite and chert. Massive and platy dolomite.	Rietfontein and Otavi Valley Occurrences. Abenab mine
	Abenab Stage		Platy limestone Pink dolomite Platy to massive dolomite Shale Tillite	Abenab West mine
	Berg Aukas Stage	Upper	Massive dolomite	Berg Aukas mine
		Lower	Laminated dolomite Limestone Dark dolomite, shale	Nosib mine
Nosib Series			Quartzite and conglomerate Shale and ironstone	

the broad synclinal valleys between the ranges. Solution cavities and irregular hollows on the weathered surface of the dolomite are usually occupied by these surface deposits.

Structurally, the Otavi Mountain Land may be subdivided into a southern intensely folded and faulted portion and a northern gently folded one devoid of major faulting. The Tsumeb and Abenab mines are both situated in the northern part. The fold axes trend roughly east-west parallel to the strike of the Basement Complex. Overfolding and overthrusting to the north are common in the southern portion. Several smaller tear faults lie at an angle of 45° to the two large strike faults which are interpreted as overthrusts. A generalised geological map of the area is given by C. M. Schwellnus (1945).

The Abenab mining area is situated in the extreme north-east of the mountains, about 20 miles north of the town of Grootfontein. It lies on the southern limb of a major syncline and the strata dip steeply (60°-80°) to the N.N.W. A group of beds about 2,000 ft. thick has been mapped in detail as shown on the accompanying simplified geological map of the mining area (Fig. 1). The main structural features are:*

*These data, and the geological descriptions in the following chapters, are based on information contained in private reports by the geological staff of the S.W.A. Company.

(1) *The Abenab fault.* This is a single large dislocation causing the massive dolomite-limestone contact to transgress across the platy limestone. Drag phenomena are prominent along this major zone of movement which is thought to persist over a distance of several miles to the E.N.E. and W.S.W. Tectonic breccias with strong shearing effects and no vanadium mineralisation occasionally mark the fault line. Younger than these and situated on and marginal to the same fault line, occur the pipe-like breccia bodies, circular in ground-plan, with which is associated the bulk of the vanadium mineralisation as exemplified by the old Abenab deposit.

(2) *The Abenab West disturbed zone.* This is a prominent zone of plastic deformation and associated faulting extending across the Abenab West mining area more or less at a position coinciding with the pink dolomite horizon. Here the deforming stresses were relieved along a number of "pinches" and "swells" caused by pairs of short arcuate faults lying in the same strike direction. The intensity of the faulting decreases rapidly downwards and on the 340 ft. level they have died out or are strictly dip-slip features. The lead, zinc and vanadium ores of the Abenab West mine are directly associated with these disturbances. The ore body is also affected by one or more "rolls" or subsidiary folds within the synclinal limb. It is possible that these "rolls" localised the deformation and associated mineralisation which are found to diminish both in depth and laterally away from the crest of the "roll" between the 150 ft. and 340 ft. levels. However, there are indications that the arcuate faults are younger than the emplacement of the sulphides. It may be suggested that they are due to pressure relief at the present surface and therefore relatively recent.

IV THE ABENAB WEST ORE BODY

Description

The valuable minerals of the Abenab West mine are found disseminated in a red-brown unconsolidated clay known as the ore body. It has an elongated irregular groundplan with constrictions and enlargements determined by the pairs of arcuate faults along the disturbed zone. The mineralised clay and associated Zinc Reef have been followed over an intermittent outcrop distance of 2,400 ft. The ore body varies in width from 2 to 20 ft. and persists to a depth of at least 450 ft. The mine has five levels at intervals of 50 ft. down to a depth of 250 ft. and a sixth level is situated at 340 ft. below the surface. An opencast about 50 ft. deep marks the original outcrop of the mineralised clay. Because the initial mining operations were more or less exploratory in nature, the main drives of the upper five levels were excavated in the ore body itself. Owing to the unconsolidated nature of the mineralised clay, this procedure resulted in extensive caving and collapse during the rainy seasons. Fortunately, assay sampling and geological mapping had been completed in time, but large portions of the 100 ft., 150 ft. and 250 ft. levels were missing or inaccessible when the samples for the present study were collected.

A convenient system of reference in use at the mine will also be employed here to specify localities. The levels are more or less linear in form and are simply divided into compartments 5 meters (16·4 ft.) in width, and are numbered east and west from the old Inclined Shaft; 340 ft. EO1, for example, is the first compartment east of the reference line on the 340 ft. level.

The form of the ore body is best illustrated by transverse sections (Fig. 2) showing the stratigraphic relations and structural control already described. The

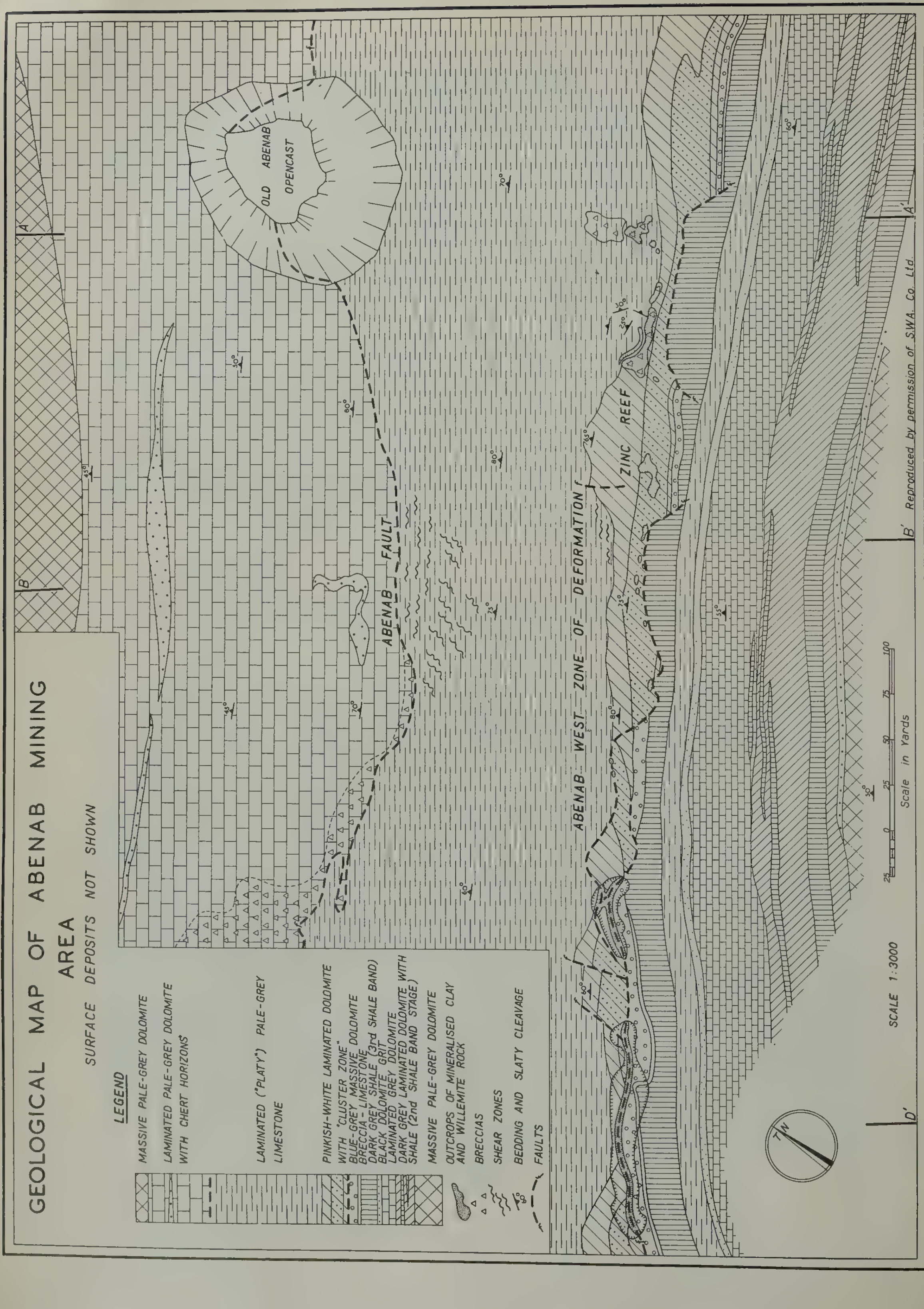
GEOLOGICAL MAP OF ABENAB MINING AREA
SURFACE DEPOSITS NOT SHOWN
LEGEND
MASSIVE PALE-GREY DOLOMITE
LAMINATED PALE-GREY DOLOMITE WITH CHERT HORIZONS
LAMINATED ("PLATY") PALE-GREY LIMESTONE
PINKISH-WHITE LAMINATED DOLOMITE WITH "CLUSTER ZONE"
BLUE-GREY MASSIVE DOLOMITE
BRECCIA - LIMESTONE
DARK GREY SHALE (3rd SHALE BAND)
BLACK DOLOMITE GRIT
LAMINATED GREY DOLOMITE
DARK GREY LAMINATED DOLOMITE WITH SHALE (2nd SHALE BAND STAGE)
MASSIVE PALE-GREY DOLOMITE
OUTCROPS OF MINERALISED CLAY AND WILLEMITE ROCK
BRECCIAS
SHEAR ZONES
BEDDING AND SLATY CLEAVAGE
FAULTS
OLD ABENAB OPENCAST
ABENAB FAULT
ABENAB WEST ZONE OF DEFORMATION
ZINC REEF
A
A'
B
B'
D'
SCALE 1:3000
Scale in Yards
Reproduced by permission of S.W.A. Co. Ltd.

FIG. 1.

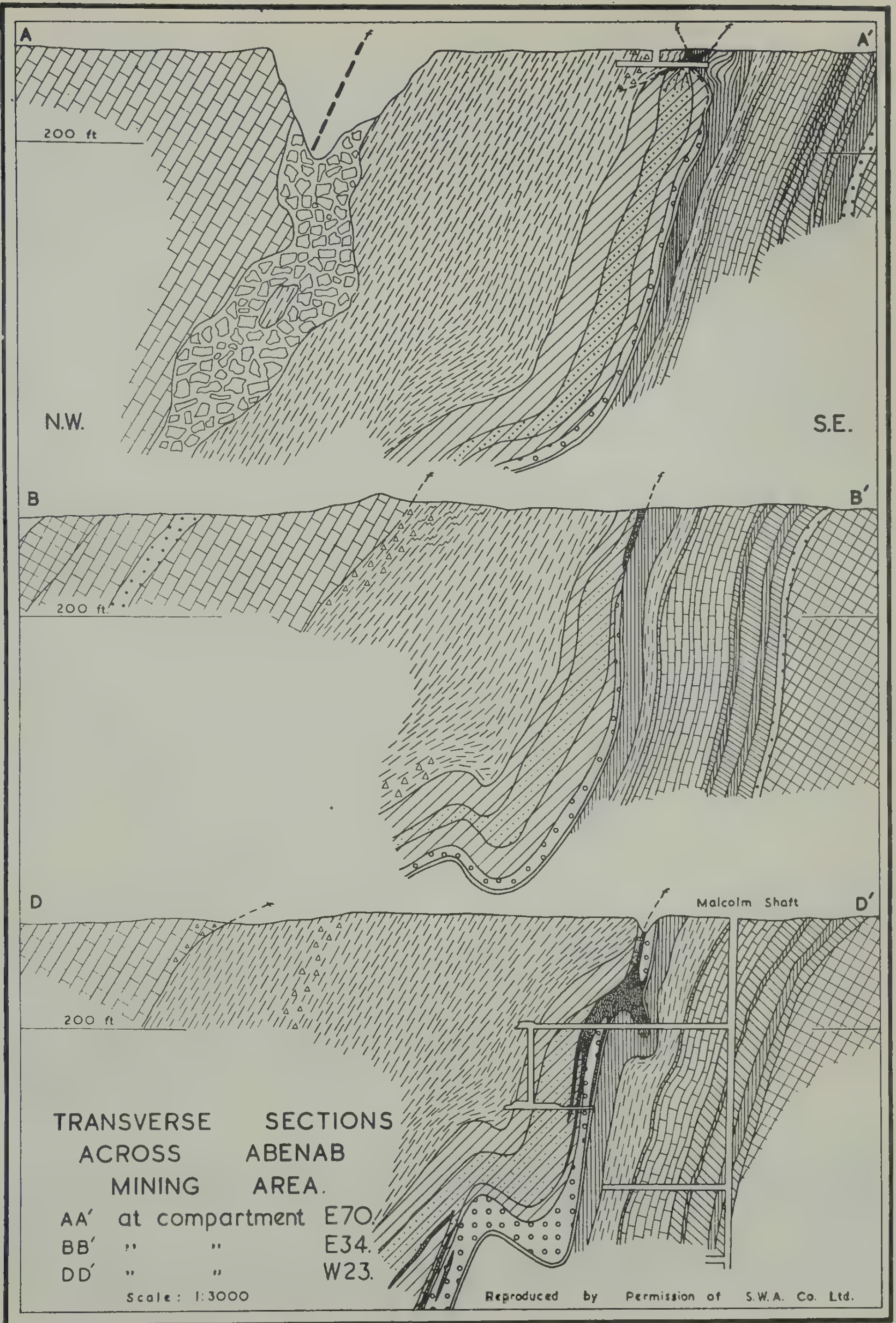
A
A'
200 ft
N.W.
S.E.
B
B'
200 ft.
D
D'
Malcolm Shaft
200 ft
TRANSVERSE SECTIONS
ACROSS ABENAB
MINING AREA.
AA' at compartment E70.
BB' " " E34.
DD' " " W23.
Scale: 1:3000
Reproduced by Permission of S.W.A. Co. Ltd.

FIG. 2.

thickest and richest part of the mineralised clay occurs along the crest of a "roll" plunging from shallow E.N.E. to moderate W.S.W. between the 150 ft. and 340 ft. levels. The ore body becomes narrower in depth and less irregular in plan. It does not always form one continuous body, but isolated stringers of clay with lesser mineralisation may occur in the adjacent disturbed dolomite. On the 340 ft. level a clay band about 1 ft. thick is found on the shale contact 20-30 ft. south of the main ore body. A longitudinal projection of the ore body (see Figs. 7 to 12) has indefinite boundaries where the ore content of the clay gradually diminishes or the clay passes into barren patches of brecciated country rock.

Bedded cave deposits: Filled caves a few feet in diameter are common features in the clay of the ore body. They usually contain a horizontally layered sequence of well-sorted coarse and fine material. The finest layers consist of an impalpable reddish clay which may be thinly laminated. These layers often grade downwards in, or are sharply separated from coarser gritty material with the following constituents: quartz, zircon, epidote, magnetite and other detrital minerals, red and black limonitic lumps and fragments of cerussite, descloizite and vanadinite that may be slightly abraded. The composition of the coarser layers varies from place to place: while they are dominantly arenaceous in one deposit, they may consist almost wholly of black oxides or of cerussite in another. Transported fragments of the vanadium minerals are minor constituents. The clay and gritty layers alternate with each other and with still coarser rubble containing country rock fragments and limonitic lumps

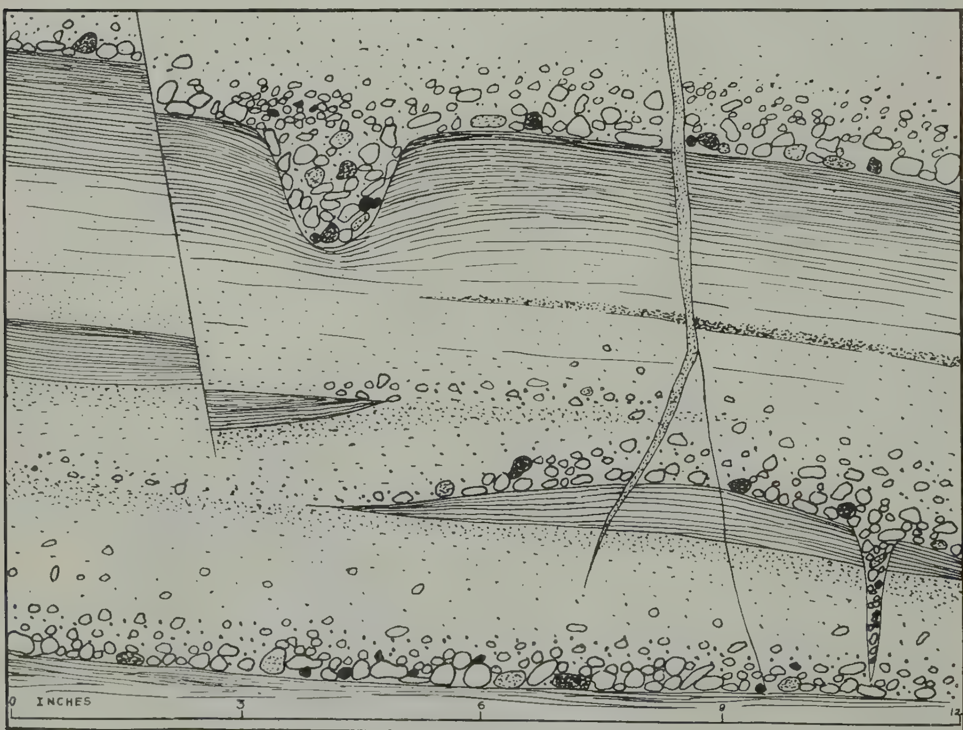

FIG. 3. Bedded Cave deposit on the 340 ft. level, A. W. mine.

up to a few centimeters in diameter. Flat pieces of platy limestone and shale are sometimes found to lie with their largest dimensions horizontal in these deposits.

A section across one of the well-preserved cave deposits at WO3 on the 340 ft. level is shown in Fig. 3. The coarsest layers consist of soft weathered shale pieces averaging 5 mm. in diameter; these grade upwards into sand and laminated clay which is sharply separated from succeeding pebble layers. It is evident that deposition of each bed started with coarse and heavy material and was terminated by mud. Nine such composite beds were counted in this deposit. The coarse layers transgress across wedge-like layers of clay, and fill fissures in the deposits below. Subsequent joints and small displacements affect the bedded material, which is seldom consolidated. However, sandy layers cemented by calcite have been found with perfect preservation of bedding planes.

On the 150 ft. level and elsewhere, such horizontal cave deposits are more than a hundred feet in diameter. It is clear that they constitute no insignificant part of the ore body and also carry considerable amounts of the ore minerals. Even where no layering is apparent, as in the stope area of the 200 ft. level, the mineralised clay contains a large proportion of rubble including sand and country rock fragments. The ore body, furthermore, often exhibits a steeply dipping banded structure as at WO1 on the 100 ft. level. Vanadinite and descloizite have crystallised *in situ* in the banded ore body, apparently as crusts conforming to the banded structure. It is suggested that these parts of the mineralised clay represent tilted cave deposits whose formation antedates the vanadium mineralisation. Slumping and deposition of bedded clay and sand by subterranean waters appear to have taken place over a long period of time. Perhaps the major part of the ore body is of this origin, while only the most recent and perfectly layered cave deposits are readily recognised.

Mineralogy

The following minerals were identified in the Abenab West clay:

galena		goethite
sphalerite	Primary	hematite
pyrite		montmorillonite
bournonite		kaolinite
cerussite		calcite
anglesite		dolomite
covellite		quartz
willemite	Secondary	muscovite
vanadinite		golden-brown mica
mimetite		tourmaline
descloizite		magnetite
		zircon
		rutile
		epidote
		kyanite
		allanite
		anatase
		microcline

Galena, PbS

Galena forms the bulk of the partly altered sulphide remnants in the iron oxide clay. The isolated blocks may be more than a hundred pounds in weight but are

usually much smaller. The central portions consist in large part of fresh granular galena, surrounded by a shell of compact or drusy cerussite and a little anglesite.

The most noteworthy features of the massive galena are its texture and structure which are brought out well by etching with Fackert's reagent (alcoholic HNO_3). While regular granitoid mosaics are sometimes found, the majority of specimens show a well-developed foliation. The flattened crystal grains are arranged in parallel layers that are plane or curved over short distances. The individuals in each layer are not uniformly orientated, as shown by their variable cleavage directions. In granitoid galena cleavage surfaces have a nearly constant size of 3-5 square mm., but in foliated varieties the texture is usually finer. The granularity may be seen to change progressively from coarse to finely crystalline in a single specimen.

The foliated galena must have developed under conditions of differential stress. When formed by free crystallisation or replacement, galena shows no tendency towards preferred orientation. Ramdohr (1950, p. 442) describes an aggregate with less pronounced foliation as typical recrystallisation texture in galena. It seems quite clear, therefore, that the Abenab West ore body was subjected to deformation during or after the deposition of the sulphides. One specimen showed an interesting structure which may be interpreted as due to folding of the ore accompanied by recrystallisation (Fig. 4*a*). A core of coarse granitoid galena is surrounded by two arcuate zones of foliated galena incorporating shreds of pyrite. The granitoid core may be explained by relief of pressure along the crest of the fold, enabling galena to recrystallise locally in undisturbed fashion.

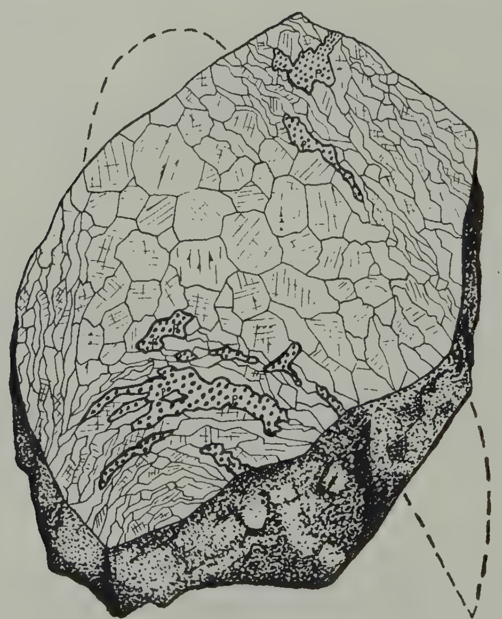

FIG. 4(*a*). Fold structure in Galena.
Galena cleavage ruled and pyrite stippled. (× 1)

Flowage and recrystallisation of galena during deformation by pressure are well-known phenomena and have been described in ore from Rammelsberg, Germany (Lindgren and Irving, 1911; Ramdohr), Slocan, British Columbia (Uglow, 1917) and Coeur d'Alene, Idaho (Waldschmidt, 1925). The Abenab West ore was not as intensely deformed as the "gneissic galena" of Slocan, or the "Bleischweif of Rammelsberg, both being areas of dynamic metamorphism. At Abenab West the galena shows no deformation twinning or translation gliding; single crystals are not excessively drawn out, sphalerite and pyrite are not crushed and there is no flowage or "augen structure" around the harder grains. The indications are more those of recrystallisation under local stress conditions, perhaps connected with shearing. The finer-grained foliated galena was probably formed where the stresses reached a maximum.

It is unfortunate that the remnants of primary ore are too few and far between to enable direct correlation between the foliation of the galena and the major structures. Evidence of subsequent disturbances in the bedded clay more-over indicates that the sulphide remnants are not necessarily preserved in their original position. It is significant that the form of the ore body shows the effects of both folding and faulting.

Sphalerite, ZnS

Remnants of sphalerite comparable to the large blocks of galena are not found at all in the oxidized ore body. The only sphalerite encountered was scattered grains in the galena, usually not discernible in hand specimens. These grains consist of polygonal aggregates with polisynthetic twinning lamellae and no obvious signs of deformation.

Pyrite, FeS_2

Small round grains and interlocking aggregates of finely crystalline pyrite are scattered in galena. In foliated ore, blebs of pyrite are strung out parallel to the foliation planes. Minute inclusions of pyrite also occur in sphalerite.

Bournonite, $CuPbSbS_3$

Microscopic inclusions of bournonite occur in the galena. Its reflective power is so near that of galena that its presence was not suspected until galena was subjected to structure etching with alcoholic nitric acid, which left the inclusions unattacked. The mineral was found to be negative towards the standard etch reagents, is distinctly anisotropic, is harder than galena and has a slightly lower reflectivity with a greenish tinge. The grains are equidimensional in form and range from a few microns to about 0·1 mm. in diameter. They occur especially along the contacts between galena crystals or moulded on pyrite, but also as idiomorphic lath-like crystals within galena. Several of the larger grains showed admirable polisynthetic twinning in one direction, thus confirming the identification as bournonite. The presence of the twinning lamellae, which are perfectly straight and undeformed, revealed the weak reflection pleochroism which could not otherwise be recognised.

Tennantite, $(Cu, Fe)_{12}\ As_4S_{13}$

This mineral was not encountered in the sections studied, but its presence may be suspected as it occurs in the sulphide ore of drill hole No. 23 in exactly the same way as bournonite in the ore under consideration.

Cerussite, $PbCO_3$

Cerussite is the predominant oxidation product of galena in the Abenab West

ore body. It occurs (*a*) as direct replacement and as cavity linings around remnants of the sulphide, and (*b*) as disseminated crystals, crystal aggregates and transported fragments in the ferruginous clay. When pure, the mineral is perfectly colourless and transparent with a resinous lustre. In the immediate vicinity of galena it usually becomes black owing to innumerable specks of unreplaced galena, the identitiy of which was proved by means of the ore microscope. The massive cerussite consists of an irregular aggregate of colourless, bluish, grey and black patches representing various stages in the total oxidation of galena. Disseminated crystals are often opaque white or grey and colour variations within a single crystal are common.

The contact between cerussite and galena is usually convex towards cerussite and is delicately serrated. The replacement proceeds across the foliation in galena, but the form of the contact is sometimes related to the curvature of the foliation planes. Veinlets of cerussite are seen to follow grain boundaries and cubic cleavage directions in galena and a striking grid-like pattern may occasionally result.

Cerussite crystals deposited at some distance from the source of the lead are found as nests and clusters in the ore body. Individuals are usually a few millimeters in diameter, but tabular crystals measuring several centimeters in length have also been found. They show pinacoidal and subordinate pyramid faces and are nearly always striated in the zone of the a-axis. The crystals are typically grouped as interpenetrating aggregates and twins.

Anglesite, $PbSO_4$

Anglesite is intimately associated with massive cerussite around remnants of galena. The mutual boundaries of cerussite and anglesite give no evidence of diversity in age and both may present idiomorphic outlines. One specimen of oxidized lead ore showed a cavity exclusively occupied by anglesite in the form of transparent equidimensional prisms ranging from 0·5 to 5·0 mm. in diameter. The mineral is undoubtedly very subordinate to cerussite in the Abenab West mine.

As both anglesite and cerussite are colourless to grey and have similar high refractive indices, it was found impossible to distinguish them in hand specimens of massive lead ore, even with the aid of the binocular microscope. Under crossed nicols anglesite is easily recognised by its relatively low birefrigence, while fragments and thin sections of anglesite show a much more distinct cleavage than cerussite.

Staining: A useful method of distinguishing between cerussite and anglesite without the aid of polarised light is provided by staining them with potassium chromate solution. Cerussite appears to adsorb chromate ions much more readily than anglesite. By immersing the crushed minerals for two or three minutes in a hot concentrated solution of K_2CrO_4, cerussite is stained a transparent lemon yellow colour (rather like descloizite) while anglesite remains colourless. The grains are best observed in transmitted light but the difference is also visible in reflected light.

The following minerals are unaffected by this treatment but are not liable to be confused with anglesite: descloizite, vanadinite, willemite, smithsonite, siderite, dolomite and calcite. If it is desired to indicate the presence of calcite also, the method may be combined with Lemberg's procedure for distinguishing between calcite, aragonite and dolomite by first immersing the grains in a 10% solution of silver nitrate. Calcite receives a coating of reddish brown silver chromate. (Twenhofel and Tyler, 1941, p. 130.)

Covellite, CuS

In polished sections of the partly oxidized galena, covellite is nearly always

present as shreds and blebs of microscopic size enclosed in cerussite. Cerussite is locally riddled with covellite laths in random orientation. Veinlets of covellite also fill cracks in sphalerite. The covellite has normal optical properties and there is little doubt that it is of secondary origin. According to Ramdohr (1950, p. 449), covellite occurring as above is typically derived from small quantities of copper present in inclusions of tetrahedrite-tennantite and bournonite, or from copper occurring in solid solution in galena itself.

Willemite, Zn_2SiO_4

Except in the "Zinc Reef" where it forms the bulk of the ore, willemite is a comparatively rare and unimportant mineral in the Abenab West mine. It has been found on all levels but only near the eastern margins of the ore body. There is no difference in properties between the willemite in the mine and in the Zinc Reef, which is really a continuation of the ore body itself.

Drusy colourless willemite encrusts earlier idiomorphic cerussite on the 150 ft. level and thin white veins of willemite also traverse massive cerussite. In bedded cave deposits a few fragments of willemite were found in cerussite-rich layers. Willemite occasionally lines small cavities in botryoidal limonite.

Willemite rock, with its characteristic red iron oxide colouration occurs as fragments cemented by calcite on the 340 ft. level. Here the willemite has been partly replaced by calcite, as shown in Plate IV, C, D and E. The zonal structure of willemite, which is otherwise hardly discernible, is very well revealed by the replacement process. Ghost-like dust relics, preserving basal and prismatic sections of willemite, are found in the vicinity of partly replaced remnants. Perfectly fresh willemite in calcite is also encountered. A hollow limonite cast with the form of willemite, which was also found on the 340 ft. level, indicates that some willemite has been dissolved and removed.

Vanadinite, $Pb_5Cl(VO_4)_3$

This mineral is an important constituent of Abenab West ore and its mode of occurrence in this mine seems to be somewhat exceptional. Typical Abenab West vanadinite consists of transparent brownish-yellow hexagonal prisms terminated at one end by steep pyramidal faces. The largest crystal observed measured $1 \cdot 3 \times 0 \cdot 4$ mm. and the smallest crystals are probably of sub-microscopic dimensions. These prisms are often welded together at one end to form radiating bunches and spheroids. In some parts of the mine, where vanadinite forms the bulk of the ore, such aggregates project from central ribs of massive crystalline vanadinite into small cavities on either side. In this manner drusy masses consisting almost entirely of vanadinite are formed, the only contamination being ferruginous material. Red and brown ferric hydrates were incorporated by massive vanadinite during crystallisation, while yellow-limonite is found to coat the crystal aggregates. Black layers containing iron and manganese are sometimes found on terminal faces of vanadinite. The earlier ferric hydrates are so intimately associated with the massive vanadinite that they would be difficult if not impossible to separate mechanically.

On the 200 ft. level at W31 vanadinite was found, differing in habit from the above. The crystals are up to 2 mm. in length, steel-grey in colour and barrel-shaped with terraced prism faces due to parallel growth. Minute crystals of descloizite are deposited on this vanadinite, which is enclosed by mud and rubble. Smaller vanadinite crystals in small drusy cavities next to these, are smooth yellowish hexagonal prisms terminated by the basal pinacoid instead of the pyramid.

Vanadinite is also found on porous varieties of hard brown limonite where minute, practically colourless prisms form isolated spheroids and radiating bunches in the cavities. The crystals are less than 0·2 mm. in length and the average diameter of the bunches is 0·4 mm. Isolated prisms of the same dimensions are implanted on limonite while infinitely smaller ones from 0·02 to 0·08 mm. in length, lie at random and form a felt-like mat. These tiny crystals sometimes occur in considerable quantities as a porous, greyish white mass that looks earthy rather than crystalline. A few grams of such material were found deposited together with quartz crystals on a weathered surface of massive cerussite surrounding galena.

Vanadinite is sometimes enclosed by calcite of later crystallisation which partly fills drusy cavities. Some difficulty may be experienced in separating this vanadinite from calcite, but there is no reason to believe that large amounts of the ore occur in this way.

In crushed ore samples single prisms of vanadinite usually occur in addition to aggregate fragments showing a sub-parallel arrangement of crystals. Two or more crystals that are attached on their prism faces are not always sub-parallel but may cross each other in any direction. Crystals may exhibit a tapering form due to early interference during radial growth. Small prisms are sometimes seen to project at various angles from the prism faces of larger ones. No doubly terminated crystals of vanadinite were encountered during this investigation.

The pyramidal faces are often curved and are invariably dull owing to a frosted, uneven surface. There is no great variety of crystal forms; the steep third order pyramid $\{21\bar{3}1\}$ and the corresponding second order pyramid $\{11\bar{2}1\}$ appear to be dominant, while the basal pinacoid $\{0001\}$ is rare and does not occur in combination with the pyramids. No cleavage could be seen when vanadinite crystals were crushed, the fracture being sub-conchoidal and hardness = 3. Within the transparent prisms zonal traces parallel to the terminal faces were often observed. These planes, which have no influence on fracturing, are interpreted as the result of crystal growth by layer formation. The fact that this type of vanadinite is always perfectly homogeneous in colour, while slight differences may be noted between specimens from different parts of the mine, shows that no changes in composition took place during the deposition of individual crystals. This condition is not unexpected in the case of small crystals and seems to apply also in the case of finely crystalline descloizite which is not zonal.

Abenab West vanadinite is not deeply coloured. Pleochroism is visible in relatively large crystals and is slightly variable, eg.:

Specimen 1: ω = brownish yellow and ϵ = yellowish brown.

Absorption $\omega > \epsilon$

Specimen 2: ω = olive brown and ϵ = reddish brown.

Absorption $\omega = \epsilon$

Small crystals are yellowish buff or may be almost colourless. Massive vanadinite is steel-grey to black in bulk, but fragments show the same brownish yellow colour and pleochroism as prismatic crystals. The lustre is from vitreous (on crystal faces) to resinous (on fractured surfaces). Abenab West vanadinite is optically uniaxial and negative in contrast with the Abenab type, which is slightly biaxial. This difference, and the lack of colour as well, is probably due to the absence of arsenic in the Abenab West mineral, as shown by comparing the following analyses:

The two varieties also differ in crystal size, crystal form, zoning and probably in paragenesis.

Vanadinite of the Abenab West kind was encountered only on and above the 200 ft. level of the mine. However, on the 340 ft. level at W24, an interesting occurrence

TABLE 2: ANALYSES OF VANADINITE

	PbO	V_2O_5	As_2O_5	Cl	H_2O-	Sum	Cl-O	Total	Analyst
Vanadinite (Abenab West)	78·65	19·01	0·0	2·42	0·22	100·30	0·55	99·75	C. J. Liebenberg (Division of Chemical Services, Pretoria).
Vanadinite (Abenab)	77·75	15·39	2·91	2·20	0·03	98·28	0·50	97·78	E. H. Strickland (Abenab Assay Office).

of a different type was discovered. Here light yellow, coarsely crystalline vanadinite forms an aggregate of short hexagonal prisms which are up to 5 mm. in diameter, bound together and partly replaced by descloizite (Fig. 4 (*b*). The replacement

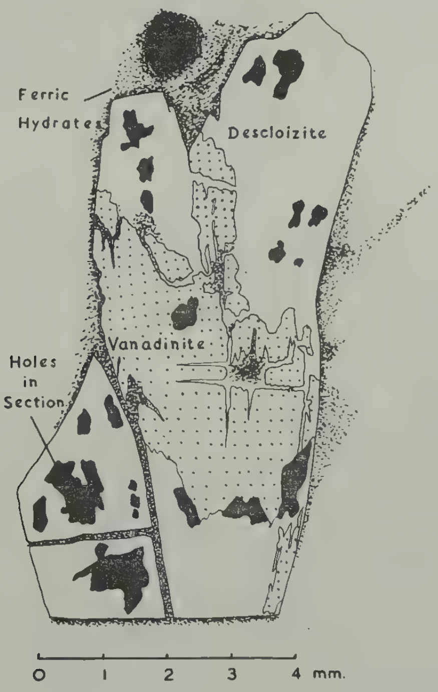

FIG. 4(*b*). Replacement of Vanadinite by Descloizite on the 340 ft. level, A.W. mine.

proceeded inwards from the prism boundaries and was guided by irregular cracks in the vanadinite. The contacts are ragged and sharp. It seems that central remnants of vanadinite were easily weathered out, leaving hexagonal cavities which may be lined with descloizite crystals. From the vanadinite that has been preserved, however, it is clear that replacement was the dominant process and not cavity filling. Complete pseudomorphs of descloizite after vanadinite occur at this locality. This vanadinite bears no resemblance to the Abenab West type already described and the occurrence is rather similar to the calcite-vanadinite-descloizite aggregates in the mineralised breccia-limestone at Okarundu. Though the vanadinite from both localities is paler in colour than the Abenab mineral, all other characteristics show that they are similar.

Further evidence of the former presence of relatively coarse vanadinite in the Abenab West mine is provided by the hexagonal casts in limonite which are described below.

Mimetite, $Pb_5\ Cl(AsO_4)_3$

As in the case of vanadinite, two types of the isomorphous arsenate, mimetite, were found:

(1) Light greenish hexagonal prisms several millimeters in length are enclosed by brown colloform limonite-goethite, associated with hexagonal mirror-like casts. This type of mimetite occurred in limonite lumps from the 100 ft. and 150 ft. levels. Concentric zones are present in this mimetite as in the vanadinite from Abenab and Okarundu. Cavities parallel to these zones are partly filled with goethite.

(2) Small white idiomorphic crystals of stout prismatic habit up to 1 mm. in length are situated in the pores of black limonitic lumps. They occur singly or in small clusters and are sometimes completely enveloped by later cerussite that fills cavities in the limonitic mass. Some of these mimetite crystals have well-developed pyramidal faces, but the stout prisms are more usually terminated by an uneven base, suggestive of parallel bunches rather than single crystals.

Under the microscope the mimetite is distinguished from accompanying cerussite by its much lower birefrigence, by slightly lower refractive indices and by the presence of hexagonal basal sections. All specimens suspected to be mimetite were identified spectrographically or microchemically. It is not an abundant mineral in the Abenab West mine but small quantities are probably present at many localities on all levels.

Descloizite, $Pb(Zn, Cu)\ OHVO_4$

Descloizite is economically the most important ore mineral in the Abenab West mine. It occurs in a variety of forms with varying crystallinity.

(1) Earthy descloizite of dirty greenish-yellow colour occurs as local concentrations in the mineralised clay and is often associated with porous, labyrinthine limonite. In samples of crushed ore it is found as thin coatings on limonite fragments and as soft yellow lumps which can be distinguished from earthy limonite by its greenish tinge in reflected light. Under the microscope and in transmitted light these grains seem opaque and amorphous, but their sub-microscopic crystallinity is revealed by a twinkling under crossed nicols. It is probable that a large proportion of this descloizite is reduced to powder during crushing. Its relative importance in the mine is difficult to judge though it does seem to be of localised occurrence and subordinate to the better crystalline varieties.

(2) Minute prismatic crystals are distributed in the unconsolidated clay, singly

or in clusters especially on the 100 ft. and 200 ft. levels of the mine. These crystals often resemble stout hexagonal prisms and with regard to shape and crystal form, they may easily be confused with vanadinite, except under high magnification, when their orthorhombic habit becomes recognisable. Some are cigar-shaped and others show tapering points; diverging aggregates also occur. The lemon-yellow colour is distinctive and the crystals are unlike vanadinite, usually doubly terminated. They range in size from 0·02 to 0·08 mm. in length and from 0·01 to 0·03 mm. across. This descloizite should give no difficulty during concentration as it can be liberated without crushing by simple agitation in water. It is abundant in the patches of red iron oxide.

Small elongated and equidimensional descloizite crystals are sometimes deposited in the cavities of porous stalactitic limonite. They occur singly and may be doubly terminated when flat-lying, but usually project from the surface of the limonite at various angles. These crystals are easily detached by crushing. Descloizite was never found enclosed in limonite and is evidently later in the paragenetic sequence.

(3) Megascopic crystalline descloizite is restricted to two or three rich portions of the ore body. The best specimens were found at W54 on the 340 ft. level, where drusy aggregates completely enveloped by wet ferruginous mud, attain a few centimeters in length. These aggregates consist of pyramidal crystals piled up along their crystallographic a-axes (according to the new orientation based on X-ray crystallography, see W. E. Richmond, 1940) to form bunches which are arranged in subparallel, radial or branching groups. Aggregates and fragments lie at random in the clay, and remnants of porous iron oxide from which crystallisation had started are sometimes attached to their bases. The habit and coarse crystallisation clearly indicate that this descloizite has grown in open cavities. The crystal edges are perfectly sharp and show no effects of transportation, except in one specimen that was found to be well-rounded probably by a strong water current which had passed along the cavity wall. It is concluded that the drusy descloizite was dislodged by caving mud, a process which may have been connected with a rising ground water level in the mineralised zone at some remote period.

Similar but smaller crystal clusters up to half a centimeter in height project from thin layers of compact descloizite found to alternate with steeply dipping bands of red iron oxide and clay, also on the 340 ft. level where the ore body is only 2 ft. thick at W27.

The descloizite aggregates described above are similar in habit to the much coarser aggregates which were often encountered in open cavities in the old Abenab breccia body. They differ from the latter in that the crystal faces, though terraced by parallel growths, are smooth and lustrous and lack the dull black exterior characteristic of most coarse descloizite in the Otavi Mountain Land.

Between the 150 ft. and 200 ft. levels in the Abenab West mine, an extremely rich pocket consisting of several tons of nearly pure descloizite was encountered. The whole mass is surrounded by ferruginous clay, which is above the local water table and dry. This descloizite was deposited *in situ* and shows no signs of disturbance. It appears that crystallisation commenced in and around blebs of red and yellow oxides which are now riddled by finely crystalline black descloizite. This black variety, which shows smooth pyramidal faces a fraction of a millimeter in diameter, is seen under the microscope to consist of pale yellow descloizite with a diffuse and irregular black staining. It is intimately associated with coarser crystal aggregates having a green to greenish-brown colour. These grade into massive crusts of descloizite which are surmounted by drusy, brown composite clusters similar to those described from the 340 ft. level. The brownish colour of these crystals is due to the presence of orange-red

pleochroic zones in a lemon-yellow groundmass near the outer surfaces of the crystals. They grew in open cavities that were subsequently filled with calcite, but this calcite has since yielded to weathering influences and only the delicate rosette-like shells described elsewhere are left behind. The whole aggregate of weathered calcite, black granular descloizite and drusy crystal aggregates forms a friable, porous mass.

In samples of ore and concentrates, coarsely crystalline descloizite is present as translucent crystal fragments in various tints of yellow and brown, with zonal streaks in orange and black, and often with indefinite black staining in patches.

(4) Descloizite in calcite: Contemporaneous intergrowths of descloizite and calcite occur on the lower eastern levels of the mine. Fragments of willemite rock and dolomite country rock, strongly infiltrated by ferruginous material, are cemented by transparent white calcite. Thin crusts of lemon-yellow crystalline descloizite coat the breccia blocks and small crystal clusters are sparsely distributed in the calcite. The descloizite content of this ore is lower than that of the typical Abenab breccia, but it is similar in that some descloizite remains attached to rock fragments and calcite after crushing.

On the 250 ft. level orange-red descloizite is embedded in calcite in similar fashion, but instead of encrusting rock fragments, it projects from intricate networks of black microcrystalline descloizite that were formed in earthy limonite. This is essentially the same type of ore as that from the rich pocket above the 200 ft. level, where weathering above the water table has almost freed the descloizite from calcite.

(5) Compact crusts of descloizite coat country rock, calcite, hard brown limonite and even dry ferruginous clay. They range from a fraction of a millimeter to several millimeters in thickness and vary in colour from light yellow to black though green is the usual colour. These descloizite crusts often have a crystalline surface but completely smooth layers are also found. They are typically of botryoidal appearance even when showing crystal faces. Both amorphous and crystalline layers, growing in fantastic forms around calcite and projecting points of wall rock, were seen to encrust the walls of a water channel on the 340 ft. level at the contact of the ore body and the Pink Dolomite. Small solution cavities and fissures in the dolomite adjacent to the ore body often carry thin botryoidal crusts of black descloizite, while similar layers are found to traverse the clay and line cavities therein. A thin veneer of light yellow and greenish descloizite is commonly found on massive and pipe-like limonite. A brecciated mass of limonite rods, red spheroidal iron hydrates and lumps of yellow ochre was found with each fragment neatly enveloped and bound together by compact descloizite. Coarse yellow vanadinite crystals are similarly cemented by descloizite on the 340 ft. level.

Except where minute layers occur in crevices of the country rock, this type of descloizite is of considerable economic importance. Some difficulty may be expected in separating it from hard brown limonite and from country rock.

(6) Descloizite pseudomorphs after vanadinite: A hexagonal prism 1·5 mm. long, built of finely crystalline descloizite, terminated by a steep pyramid and covered by brownish descloizite crystals, was encountered in a sample from the 340 ft. level. When broken, it was found to contain a central cavity lined with minute green descloizite crystals. It was obviously a pseudomorph of descloizite after vanadinite, closely related to the partly replaced vanadinite of the 340 ft. level already described.

More striking examples of pseudomorphism were found near the Zinc Reef, at E55 on the 50 ft. level. Grass-green hexagonal prisms with an average diameter of half a centimeter were arranged cross-wise and in sub-parallel aggregates like the coarser cherry-red vanadinite of Abenab. Pyramidal faces could be recognised, but

the whole aggregate was firmly enclosed by brown clay hardened by calcite and limonite. Under the microscope each prism was seen to consist of a crystalline mesh of acicular, almost opaque green crystals bound together by calcite. Spectrographic and microchemical tests proved that the green mineral is cuprian descloizite containing roughly equal proportions of zinc and copper. This is the highest copper content found in descloizite of the Abenab area. In transverse sections the calcite could be seen to fill fractures and concentric zones in each prism. This is clear evidence of the characteristic zoning of vanadinite of the Abenab type, for these zones are emphasized by weathering and the fractures filled by calcite in numerous specimens from Okarundu.

Pseudomorphs of descloizite after vanadinite are mentioned in Dana's System of Mineralogy (Palache, Berman and Frondel, 1951) while the replacement of vanadinite by descloizite is illustrated by means of chromatographic contact prints in a paper by Williams and Nakhla (1951).

Optically and chemically, descloizite from South West Africa is not homogeneous. In Abenab West descloizite, a lemon-yellow component is dominant. The smaller crystals may appear to be uniformly yellow, but under the microscope they are seen to contain thin zones of an orange-red pleochroic component. The larger crystal groups usually become progressively darker from a yellow-orange base to a greenish black or lustrous black terminal pyramid. In thin section the reason is found to be the presence of thin black zones that increase in abundance towards the tip of the crystal and which are closer together along the faces than in the interior. An analysis of such optically inhomogeneous material from W54, 340 ft. level, is presented in Table 3. The striking colour variations in descloizite are caused by small differences in composition which are further discussed in a later section. A consideration of the analyses of nearly pure components from other localities proves that the modifying influence of the red and black zones in Abenab West descloizite is negligible. Though other specimens may contain different proportions of the various components and are therefore found to differ slightly in chemical composition, the analysis quoted above may be taken to represent Abenab West descloizite in general. Locally more cupriferous varieties have been found, as in the Zinc Reef.

Mineral A

Unidentified orange-red rhombohedra about $0 \cdot 02$ mm. in diameter, have been deposited on limonite and on prismatic descloizite in the cavities of rod-like lumps. They dissolve in HCl, yield a positive test for iron and do not fuse in an alcohol flame. The powdered mineral is translucent but deeply orange in colour and non-pleochroic. It is birefringent and has refractive indices $> 2 \cdot 0$. The mineral was found in only one specimen from the 340 ft. level.

Hydrated ferric oxides

The reddish brown Abenab West "clay" consists to a large extent of hydrous iron oxide mixed with subordinate amounts of clay minerals. Analyses of representative samples (See Table 4) show that the Fe_2O_3-content of the ore ranges from about 10 to 50%. Probably most of this is in an earthy form, but local concentrations of hard and compact limonite, mostly deposited in the colloidal state, are common features of the ore body. Posnjak and Merwin (1919) have shown that there is no mineral species "limonite" with a formula of $2Fe_2O_3.3H_2O$ and that $Fe_2O_3.H_2O$ (goethite) is the only true hydrate of ferric oxide. The massive, meta-colloidal and earthy materials for which the general term "limonite" is employed in this paper,

TABLE 3: ANAYLSES OF DESCLOIZITE

* New analyses.

	1.*	2.	3.*	4.*	5.*	6.
PbO	55·21	55·89	54·55	54·78	57·40	52·87
ZnO	18·89	19·12	18·01	19·19	17·66	21·65
CuO	1·05	1·07	1·76	1·19	0·63	0·04
V_2O_5	21·28	21·54	20·82	21·00	21·39	21·35
As_2O_5 ...	0·35	0·36	0·81	0·20	0·39	1·79
H_2O- ...	nil	nil	0·04	0·03	0·03	n.d.
H_2O+ ...	2·00	2·02	2·10	2·20	2·00	2·31
Total ...	98·78	100·00	98·09	98·59	99·50	100·01

1. Greenish yellow crystal aggregates from W54, 340 ft. level, Abenab West mine. Analyst: E. H. Strickland.
2. Greenish yellow crystal aggregates from W54, 340 ft. level, Abenab West mine. (1 recalcuated to 100.) Analyst: E. H. Strickland.
3. Green descloizite from small breccia pipe at Okarundu. Analyst: E. H. Strickland.
4. Pseudostalactitic descloizite from "sand sack", Berg Aukas. Analyst: E. H. Strickland.
5. Red descloizite from old Abenab mine. Analyst: E. H. Strickland.
6. Descloizite from Abenab. Analysts: E. Dittler and H. Hüber (1931).

consist of intimate mixtures of goethite and hematite with varying amounts of adsorbed water. Both goethite and hematite have been identified by Mr. H. Heystek with X-ray diffraction methods in composite samples of Abenab West clay. Goethite has also been recognised microscopically whereas lepidocrocite and turgite, as defined by Posnjak and Merwin, have not been encountered.

Patches of brilliant red powder were suspected to consist of minium, but it was found that their lead content is due to the presence of minute disseminated crystals of descloizite, while the red material is pure iron oxide. When heated to 900°C., the red material lost 8·8% weight and changed to a deep scarlet colour. In the mine powders of all shades from scarlet to brick-red are represented, often closely associated with similar blebs of ochreous yellow material. Amorphous red and yellow ferric oxides often occur together in nature and it is probable that both are formed at ordinary temperatures by somewhat different reactions. The yellow material is apparently essentially ferric oxide monohydrate while the orange and red substances may be hematite with water in either a dissolved or an adsorbed condition, or both. (Posnjak and Merwin, 1919.)

Hard crusts and lumps of limonite, usually brown, red-brown or black in colour, occur throughout the mineralised clay and often carry appreciable amounts of ore minerals both earlier and later in the paragenetic sequence. The lumps may be structureless and compact but are more often botryoidal (at least in part) with a colloform texture and numerous cavities between adjoining spheroids. Surfaces are frequently covered with velvety layers of earthy limonite, yellow to brown in colour, which are sometimes difficult to distinguish from thin descloizite coatings occurring in the same way. Microscopically the colloform layers are seen to consist of brown opaque bands alternating with translucent orange-red ones. The red mineral is non-pleochroic, anisotropic, and has a fibrous structure with parallel extinction and

TABLE 4: ANALYSES OF ABENAB WEST MINERALS AND OF THE ZINC REEF

	1	2	3	4	5	6
PbO	4·38	1·37	1·53	0·45	0·67	0·68
ZnO	4·66	0·31	0·44	0·63	0·85	59·20
CuO	0·18	n.d.	n.d.	n.d.	n.d.	0·09
MnO ...	<0·02	n.d.	n.d.	0·05	0·05	0·08
Fe_2O_3 ...	70·96	0·03	0·02	0·43	1·15	3·28
Al_2O_3 ...	6·37	<0·01	0·02	1·91	2·08	2·17
TiO_2	n.d.	n.d.	n.d.	n.d.	n.d.	0·11
CaO	0·90	53·52	52·80	44·53	46·77	4·48
MgO ...	0·42	0·36	0·72	1·85	2·12	0·87
V_2O_5	1·48	n.d.	n.d.	0·07	0·12	n.d.
As_2O_5 ...	n.d.	n.d.	n.d.	n.d.	n.d.	0·02
P_2O_5	n.d.	n.d.	n.d.	n.d.	n.d.	0·34
CO_2	0·17	42·38	42·26	34·08	38·98	n.d.
SiO_2	3·58	n.d.	n.d.	9·28	4·02	23·75
S	n.d.	n.d.	n.d.	n.d.	n.d.	0·02
H_2O (110°C.)	0·50	0·15	0·02	0·03	0·24	n.d.
Total ...	93·62	98·13	97·80	93·31	97·05	95·09

1. Limonite rods with a little descloizite, 340 ft. level, Abenab West mine. Analyst: E. H. Strickland.
2. Plumbian calcite from 340 ft. level, Abenab West mine. Analyst: E. H. Strickland.
3. Plumbian calcite trom the Zinc Reef. Analyst: E. H. Strickland.
4. Yellow platy limestone adjacent to (5). Analyst: E. H. Strickland.
5. Red platy limestone permeated with iron oxide next to the Zinc Reef. Analyst: E. H. Strickland.
6. Composite Zinc Reef sample. Analyst: Daniel C. Griffiths & Co., London. (Published by permission of S.W.A. Co.)

Other constituents reported:

Ag	...	0·0025% (0·82 oz./2240 lb.)
Au, Cd, Bi, Ni, Cl	...	traces
Sn, Sb, Co, F ...	...	nil.

positive elongation of the fibres. These properties correspond to goethite, $Fe_2O_3.H_2O$. The opaque brown layers, which are apparently amorphous, very probably consist of cryptocrystalline goethite, though the chemical analysis cited above indicates the presence of some anhydrous ferric oxide too. Tiny implanted globules on colloform limonite consist entirely of red radiating fibres of goethite. These globules are sometimes piled up in delicate branches crossing each other haphazardly. Some brown compact lumps are also built entirely of closely spaced goethite spheroids.

Limonite rods

Among the most characteristic forms of limonite in this ore body are the closely spaced cylindrical rods that have given rise to the local name of "pipe ore". These limonite rods are related to the spheroidal and globular masses described above. They are also built of radial goethite and amorphous limonite in alternative concentric layers and they often merge into curved sheets, mammillary crusts or structureless lumps of similar material. They range from a fraction of a millimeter to about 2 mm. in diameter and may attain several centimeters in length. Usually they have a rather

uniform cross-section along their length, tapering only at the end. The rods were always found in a vertical position except where they have been brecciated and cemented by later calcite and descloizite. Aggregates of parallel limonite rods have been completely enveloped by compact descloizite crusts on the 340 ft. level, while on the upper levels small individual crystals and bunches of descloizite, vanadinite, cerussite (rarely mimetite and willemite) have been deposited in cavities of the porous masses. Subsequently calcite may have crystallised in the voids and in some instances the whole aggregate has been firmly cemented to form a calcite-limonite rock. Limonite rods are often closely associated with mirror-like casts and remnants of earlier minerals. Irregular vesicular lumps of limonite are also found, but a linear element can nearly always be recognised, indicating a similar origin to that of the regular rods.

The exact mode of formation of the cylindrical limonite is not clear, though the process was probably similar to the formation of stalactites. The close spacing of the rods, their uniform cylindrical shape, extremely slender form and frequent coalescence are somewhat difficult to explain. These features may, in part, be due to deposition from colloidal suspensions instead of ordinary solutions.

Table 4 contains an analysis of so-called "pipe ore" from the 340 ft. level. Under the microscope the specimen appeared to consist of very dark brown limonite rods with a small amount of descloizite as individual crystals in the interstices. Moisture was determined at 110°C. only. If the unaccounted 6·46% is assumed to be combined water, the percentages of the principal constituents are theoretically as follows: goethite 62·3%, hematite 15·0%, descloizite 6·9%. It is unknown in what mineralogical form the SiO_2, Al_2O_3, MgO and excess of PbO, ZnO, CuO are present.

Limonite casts

Numerous specimens show limonite moulded on earlier crystals, especially on calcite. Where the calcite has been removed, its imprint is left in the limonite, and the surfaces originally in contact with the calcite have the appearance of a highly polished metal. The outer surfaces of thin limonite layers may still imitate the form but they are dull and usually encrusted in turn by later crops of calcite. Such "mirror-like casts" of limonite are common in the ore body, owing to the removal of the earlier crystals through weathering. The limonite layers are often exceedingly thin and appear under the microscope as transparent, isotropic golden yellow flakes. However, the colour is not characteristic and has been seen to grade from yellow through brown to black in one piece of rock. Glistening cellular aggregates in which the rhombohedral shape of former calcite crystals may be recognised are formed in this way. Especially characteristic are the casts of rounded calcite lumps terminated by the numerous sharply pointed $\{40\bar{4}1\}$ rhombohedra.

Of greater interest are the occasional casts of hexagonal prisms found in limonite. They are about 1 mm. in diameter and less in depth. In samples from the 100 ft. and 50 ft. levels, some were still occupied by the original mineral: mimetite in one instance and vanadinite in the other. In another specimen, the base of a small hexagonal cavity descended towards the centre by way of three concentric steps. The only explanation for this seems to be a connection with the zonal structure of vanadinite or mimetite. The negative crystals prove that an early generation of these two minerals must at some time have been more widespread in the oxidized ore body than at present.

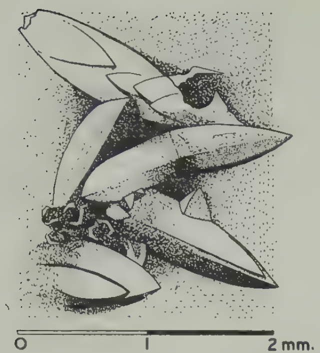

FIG. 5(*a*). Limonitic shells pseudomorphic after Calcite.

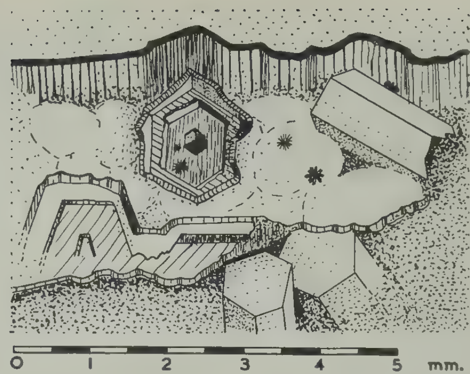

FIG. 5(*b*). Casts and shells of Limonite pseudomorphic after Vanadinite or Mimetite, and encrusted with radial groups of late Vanadinite.

Crystals that were not enclosed by limonite often received a mere film of limonitic material that has frequently survived as a hollow casing after the original mineral has been leached. Fragile shells like these are sometimes found near casts of the same mineral and they afford an even better reflection of the shape of the original crystal or crystal aggregate. The following types were encountered:

(*a*) Bunches of acute rhombohedra characteristic of calcite are the most abundant.
(*b*) Stout hexagonal prisms terminated by the base are ascribed to vanadinite or mimetite.
(*c*) One hexagonal prism with a flat rhombohedral termination was found in a sample from the 340 ft. level. It obviously indicates the former presence of willemite.
(*d*) A few square tabular shells may be due to cerussite—one had a series of parallel ridges on a pinacoidal face reminiscent of the parallel growths often seen on cerussite.

Limonite pseudomorphs after pyrite

The well-known limonitic cubes and pyritohedra are often found in shear zones in the Otavi Dolomite. They also occur abundantly at W47 in the subsidiary clay band on the 200 ft. level.

Clay minerals

Up to 29·94% Al_2O_3 has been reported in chemical analyses of representative ore samples from Abenab West. Alumina seems to be an ubiquitous constituent of the ore mud but is nearly always subordinate to Fe_2O_3. Except for a few greyish nodules, no approximately pure clay patches could be recognised in the mine. It is believed that the clay minerals are not segregated but occur in close association with the earthy ferric hydrates.

The mineralogical form in which the alumina occurs could not be established by ordinary chemical and optical means. However, the identification of the clay minerals was considered to be of particular importance, both from the genetic viewpoint and in view of the possibility that zinciferous clay is present. (See Chapter VIII.) The assistance of the South African Council for Scientific and Industrial Research was

sought and five samples were accordingly submitted to the National Chemical Research Laboratory for identification of the clay minerals by X-ray diffraction methods. The author is indebted to H. Heystek for the following results:

TABLE 5: IDENTIFICATION OF CLAY MINERALS

Sample	Description	Results of X-ray study
A1	Fine fraction obtained by decantation of Sample F16 (200 ft. level). Original sample contains 14·52% Al_2O_3. (Table 13, Analysis 4.)	Goethite and hematite; no sign of clay mineral.
A2	Fine fraction obtained by decantation of Sample G10 (340 ft. level). Original sample contains 17·92% Al_2O_3. (Table 13, Analysis 6.)	Montmorillonite is the principal clay mineral with indications of a little kaolinite and hydro-mica.
A3	Fine fraction containing 25·24% Al_2O_3 and unaccounted zinc; obtained by decantation of Sample D2 (100 ft. level). (Table 13, Analysis 7).	Calcite, goethite, hematite; incations of montmorillonite. The clay mineral is mainly kaolinite, probably in the form of a chamosite (hydrated iron silicate isostructural with kaolinite).
A4	A clay-like nodule surrounded by yellow limonite on the 150 ft. level.	The clay minerals are montmorillonite and a little hydromica.
A5	Shale from shale band 30 ft. from ore body on 100 ft. level. (For comparison.)	Calcite and mica with a trace of kaolinite.

From the above it appears that the clay minerals in the ore body are montmorillonite and kaolinite together with some hydromica. The problem is not wholly solved, however, for sample A1 indicates that a major portion of the alumina is not yet accounted for. It may be suggested that this alumina most probably occurs as the hydrated oxide in the form of diaspore or gibbsite, which are the final disintegration products of aluminous minerals and the principal constituents of bauxites. If this is the case, there is a possibility that V^{5+} ions may be adsorbed by the hydrated oxides as in the French bauxite deposits. In the clay minerals, again, V^{5+} may be present in substitution of Al^{3+} in the same way as in shales. (Rankama and Sahama, 1950.) Further work will be necessary to establish the presence or absence of combined or adsorbed vanadium in Abenab West clay. The analyses quoted in this work, however, do not indicate a significant excess of vanadium over the amounts of lead and zinc required to form descloizite and vanadinite.

Calcite, $CaCO_3$: Transparent, coarsely crystalline calcite has come to be regarded as an essential associate of vanadium ore in the Otavi Mountain Land. In the Abenab West mine it is by no means lacking, though limonite and ferruginous clay are more important gangue minerals. The calcite exhibits a few features worthy of note.

Reference has already been made to various modes of occurrence. The calcite cements brecciated country rock and massive willemite, occurs interstitially with descloizite aggregates and with stalactitic limonite, encloses vanadinite clusters and is encrusted by compact layers of descloizite. It also forms spheroidal aggregates around projecting ferruginous masses in the clay. These spheroids usually have spiny outer surfaces formed by terminal rhombohedra which may leave their imprint on subsequent limonite layers. Calcite younger than the spheroidal aggregates is usually deposited interstitially and is separated from them by thin shells of limonite. The spheroidal masses range from a few millimeters to several centimeters in diameter. One large aggregate with a diameter of 15 cm. was found at E14 on the 340 ft. level. The outer surface of the aggregate is covered with acute rhombohedra belonging to the form $\{40\bar{4}1\}$. Internally it is characterised by a concentric banding like stalagmite, due to the alternation of thin white opague and thicker purplish translucent layers of calcite. Under the microscope this aggregate shows elongated grains lying in rows with a radial disposition and traversed more or less perpendicularly by the banding. The opaque bands are due to included material and cavities among which a few liquid inclusions with very small contraction vacuoles could be recognised; these may not be primary. Similar banded calcite occurs in veins at various localities in the Otavi Mountain Land, but no aragonite has been found. In this connection it is of interest that both P. A. Wagner and H. Schneiderhöhn refer to calcite paramorphic after aragonite and described as "mammillary crusts up to 1 cm. in thickness deposited on projecting crystals of descloizite." (Wagner, 1923, p. 144.)

Some calcite crystals at Abenab West are zonal; they show opaque white zones containing negative crystal cavities and dusty inclusions. The zones are not equally resistant to weathering so that weathered calcite often consists of concentric rhombohedral shells, almost like lithophysae. The possibility that the zones differ slightly in composition is not excluded. H. O'Daniel (1930) describes tarnowitzite (plumbian aragonite) from Tsumeb, showing that the mineral is not a homogeneous mixed carbonate but is built of a zonal alternation of lead-rich and lead-poor layers each consisting of homogeneous, isomorphously mixed material. The tarnowitzite is irregularly distributed in masses containing calcite, plumbocalcite, copper, malachite, smithsonite and cerussite.

Plumbian and zincian calcite occur at Abenab West, but the content of heavy metals is low, as shown by the two analyses in Table 4. Specimen (*a*) with 1·37 % PbO and 0·31 % ZnO shows zonal structure and comes from the 340 ft. level. ω was determined as $1{\cdot}665 \pm {\cdot}0025$. Specimen (*b*) is pinkish calcite from the Zinc Reef with 1·53 % PbO, 0·44 % ZnO and $\omega = 1{\cdot}667 \pm {\cdot}0025$. These two specimens gave the strongest lead and zinc tests among twelve that were tested micro-chemically.

Dolomite, $CaMg(CO_3)_2$

In addition to blocks of dolomite country rock, thin crystalline crusts of dolomite are sometimes found in the Abenab West clay. It is easily distinguished from calcite by its characteristic pearly lustre.

Quartz, SiO_2

Two varieties of quartz, neither of which seems to have a genetic relationship with the mineralisation history, occur in the ore body.

(*a*) Sub-angular to well-rounded sand grains are present in every ore sample from the surface down to the 340 ft. level. They are slightly buff-coloured or colourless and have an average diameter of 0·1 - 0·2 mm. Smaller and more angular grains are

also abundant. There is no doubt that the sand has suffered abrasion and transportation, most probably by wind, and was incorporated in the oxidized ore body from above. According to H. Schneiderhöhn (1920) 59% of the eolian sand grains in the neighbourhood have a diameter between 0·05 and 0·15 mm. which agrees well with the material in the mine. This correspondence and the eolian origin of the sand are further corroborated by the presence of the detrital minerals described below.

A few sand grains were found to consist of sagenitic quartz: clear transparent quartz enclosing red acicular rutile crystals disposed in reticulated fashion.

(*b*) Quartz that crystallised *in situ* is found between W25 and W30 on and above the 100 ft. level. It forms the matrix of a hard and compact rock containing euhedral crystals of cerussite, descloizite and calcite. These minerals are not worn or brecciated and were not transported before their cementation by quartz. The mass is coloured by streaks of red and yellow iron hydrates and contains cavities lined with descloizite and quartz, the quartz being latest. The descloizite is often attached to calcite. These two minerals are finely crystalline, averaging about 0·2 mm. in diameter, and are pseudo-poikilitically enclosed by quartz. Cerussite is not found in close association with calcite and descloizite and is coarser grained (± 1·0 mm.) so that quartz is in this case interstitial.

This siliceous type of ore would give considerable difficulty during treatment, but fortunately it does not seem to be particularly abundant.

Idiomorphic quartz crystals were also found to be enclosed in cerussite and anglesite surrounding remnants of galena on the 200 ft. level, while drusy quartz on the same specimen is covered by extremely fine vanadinite.

Muscovite: Muscovite in the form of thin colourless plates, usually rounded in form, is a minor constituent of the ore mud.

Golden-brown mica: Small cleavage flakes of a yellow-brown mica are more abundant than muscovite, with which it is usually associated in all parts of the mine. The flakes are extremely thin and rather uniform in size, the larger ones invariably having rounded edges. They are found among the fine material of the stratified cave deposits in the mineralised clay. These facts show that both micas are detrital minerals which are not expected to have a genetic relationship with the ore.

The following properties of the golden-brown mica could be determined: The cleavage flakes are brittle to flexible but inelastic. Under the microscope the flakes are seen to be slightly pleochroic with β = greenish yellow and γ = yellowish brown. Absorption $\beta < \gamma$ Refractive indices and axial angle vary somewhat but the majority of flakes seem to have $\beta = \gamma = \pm 1{\cdot}655$ and $2V(-) = 40° - 45°$. Dispersion $\rho < v$ is distinct. Transverse sections, obtained after embedding cleavage flakes in a thermosetting plastic, showed moderate birefrigence and maximum absorption parallel to the cleavage traces. Owing to the strong yellow colour of the mineral, flakes of sufficient thickness show anomalous green interference tints. Efforts to obtain a percussion figure in order to determine the orientation of the axial plane were unsuccessful owing to the small size and brittleness of the flakes. On heating strongly on a metal plate the golden-brown mica exfoliates slightly, the colour darkens to a red-brown and the optical constants appear to remain unchanged.

Three independent spectrographic determinations revealed the presence of much Si, Mg, Fe, less Al, Ca, Mn, Ti and no K or V. The mineral is therefore not roscoelite or ordinary biotite. Though the relatively high refractive indices and large optic axial angle may perhaps suggest clintonite, it is regarded as more probable that the yellow-brown mica represents a weathered product of iron-rich biotite which has been leached of alkalis, but not a typical vermiculite.

Insoluble heavy constituents: The minerals described below in approximate order of abundance are accessory constituents of the ore mud. They were identified in insoluble heavy residues of ore samples from various parts of the mine. The ore samples were boiled in 1 : 1 HCl and the heavy fraction of each residue was separated with a bromoform-alcohol-mixture of specific gravity about 2·7. The heavy mineral grains have average diameters of 0·02 to 0·1 mm. and usually show some effects of abrasion but are poorly rounded in general. These minerals together constitue less than 5 % of the insoluble material, the balance being almost entirely made up of quartz.

Tourmaline is by far the most abundant heavy mineral encountered. Two divergent types are found, viz.:

(*a*) Ordinary pleochroic tourmaline, including several varieties that differ in colour and pleochroism as follows:

ϵ	ω
Light buff	brown
light buff	dirty green
brownish	dark green
greyish	blue

Green tourmaline is the dominant variety in the mine. It occurs mostly as stout little prisms, often doubly terminated by pyramidal faces showing hemimorphism. Elongated prisms and fragmentary grains with scarcely rounded edges are also found. The brown tourmaline has a similar mode of occurrence. Many crystals contain pinkish bubble-like inclusions with a low refractive index. These inclusions may attain surprising dimensions: in one case three inclusions together occupied roughly half the volume of the host. Under high magnification the bubble-like inclusions themselves are sometimes seen to contain one or more smaller stationary inclusions. The tourmaline also contains primary liquid inclusions having the form of negative crystal cavities and containing relatively large contraction vacuoles in motion.

The following optic constants of green tourmaline were determined with the aid of immersion media:

$$\epsilon = 1{\cdot}6215 \pm {\cdot}0025$$
$$\omega = 1{\cdot}6420 \pm {\cdot}0025$$

Blue tourmaline is comparatively rare and always fragmental in form.

(*b*) Colourless tourmaline. Slender colourless prisms with straight extinction, negative elongation and moderate birefringence provided quite a problem during the study of the insoluble heavy residues. The negative optical character of the mineral could be determined from a flash figure and the following values were obtained for its refractive indices (immersion method):

$$\epsilon = 1{\cdot}610 \pm {\cdot}0025$$
$$\omega = 1{\cdot}632 \pm {\cdot}0025$$

The same values were obtained during a repeat determination several months later. The identification of the mineral as tourmaline was confirmed by observing its ditrigonal symmetry under the binocular microscope, and by a qualitative spectrogram, which, though probably incomplete owing to insufficiency of material, proved the presence of Si, Mg, B, Al, Ca. From the available evidence and particularly the low refractive indices, the mineral is concluded to be practically pure dravite (Mg-tourmaline).

The crystals have a characteristic habit which differs markedly from that of the pleochroic types. They are invariably slender prismatic with an average length of 0·1 to 0·2 mm. and an average thickness of 0·02 to 0·04 mm. Prisms shorter than these are considered to be broken portions of larger crystals. Some of them have flat

pyramidal terminations but rectangular edges are more common; they may in part be due to the breakage of still longer ones with a basal parting.

This extraordinary variety of tourmaline is a common accessory constituent of most ore samples.

It is of interest that these colourless tourmaline prisms were also found in blebs of scarlet hydrous ferric oxide containing minute disseminated crystals of descloizite. The usual detrital minerals were not present.

The origin of the colourless tourmaline at Abenab West is somewhat obscure.

Magnetite is abundantly present in the form of small, opaque black grains that are weakly magnetic.

Zircon occurs as small round or elongated grains, often with recognisable crystal faces. Zonal structure is not unusual and inclusions are sometimes present.

Rutile: Apart from sagenitic intergrowths in quartz, rutile is a common constituent of the concentrates. Two distinct species were recognised:

(*a*) Yellow rutile is the most abundant type. Pleochroism is absent or weak with:

ϵ = orange yellow
ω = golden yellow

Slender and stout perfectly idiomorphic prisms are not rare, although fragments of parallel bunches, strongly striated vertically, are the usual forms. Smooth fragments of larger crystals are occasionally encountered, and knee twins have also been found.

(*b*) Red rutile shows somewhat stronger pleochroism and may be so deeply coloured that it looks opaque in transmitted light.

ϵ = dark blood red
ω = orange red

Geniculate twins and several unbroken, broadly prismatic crystals with pyramidal terminations at both ends were seen.

Rutile shows little or no evidence of wear and transportation. Its presence in the ore is held responsible for the positive titanium tests reported from Abenab Assay Office. Other possible accessories that may contribute titanium are titanite and ilmenite.

Epidote is quite common and is characterised by pleochroism from yellow green to almost colourless. The grains are somewhat rounded, have rough surfaces and often yield negative biaxial interference figures showing one optic axis due to the (001) cleavage.

Kyanite: Colourless tabular fragments with rounded edges, showing the characteristic parting and cleavage lines, an extinction angle of 30°, moderate birefringence and large negative 2V. Common.

Allanite: The conchoidal yellow pleochroic fragments of allanite are not very abundant. The following diagnostic properties could be determined on one well-preserved crystal elongated parallel to *b*: extinction straight, elongation ($\pm$), optic plane (010), optic angle large, refringence about 1·74 and birefringence roughly 0·01. Pleochroism is as follows:

α = brownish yellow.
γ = light yellow.

Absorption $\alpha < \gamma$.

Anatase: Rare, blue unworn fragments are present, displaying extreme refringence and birefringence approximately 0·05. A fragment showing prismatic crystal faces gave straight extinction, negative elongation and weak pleochroism in blue with absorption $\omega < \epsilon$.

A round grain of the same blue colour and high refringence gave a positive uniaxial interference figure which renders its identity with anatase doubtful. Schneiderhöhn, who describes only yellow and colourless anatase from the desert sand of the region, notes the presence of an optically positive "unknown blue mineral" very similar to the above but with a refractive index of only 1·6-1·7. (Schneiderhöhn, 1920, p. 280.)

Feldspar: A stray fragment of fresh microcline was found in the heavy concentrate.

V THE ABENAB WEST "ZINC REEF"

East of E25 the Abenab West mineralised zone was characterised by resistant outcrops of zinc silicate extending over a distance of 450 yards. Though continuous with the main ore body and closely related in origin, the "Zinc Reef" is conveniently described as a separate occurrence. It was first prospected in 1925 and subsequently abandoned because it bottomed on barren dolomite at a depth of 50 ft. During 1950-1952 it was developed on the 50 ft. level and in 1952 some ore was mined and treated.

The reef consisted of hard red material that is riddled with veins and drusy cavities lined by radiating aggregates of colourless acicular willemite—not hemimorphite as previously supposed. Microscopic investigation showed that almost the entire rock is made of compact radial willemite, coloured and contaminated by red iron oxide. Mineralogically it appears to differ markedly from the "Zinc Silicate shell" of Broken Hill, Northern Rhodesia, which "resembles a fine-grained cherty ironstone and consists of hemimorphite, zinc carbonate, lead carbonate and vanadium minerals diffused through jasperoid silica." (Pelletier, 1929.) According to L. J. Spencer (1927) "willemite is without doubt extremely abundant at Broken Hill, though so obscure mineralogically that it has hitherto been overlooked.

The crystallisation of willemite at Abenab West has mainly taken place in what are usually considered the hanging wall strata on the faulted contact of dolomite and limestone. Up to a few per cent of zinc is also reported towards the horizon where the Abenab West ore body is usually situated and in the laminated limestone between the Abenab and Abenab West fault zones. Field relations strongly suggest that the country rocks have been replaced by the zinc silicate, but microscopic evidence of replacement could not be obtained.

The zincification appears to have been intimately associated with the infiltration of ferruginous material, the red colouration of which is also observed along cracks, joints and bedding planes in the limestone near the Zinc Reef. Specimens of red and yellow ferruginised limestone have been analysed with the hope of establishing two stages in the replacement of limestone by zinc silicate and ferric oxide. However, the results (see Table 4) show no concomitant increase of iron and zinc.

Fragments dislodged from the Zinc Reef by weathering are not easily decomposed and tend to persist as pitted pieces, so thickly scattered in places that they form a thin scree.

Mineralogy

The following minerals were identified in the Abenab West Zinc Reef:

Ore minerals	*Gangue minerals*
Galena	Calcite

sphalerite
willemite
wulfenite (vanadian)
greenockite
malachite
cerussite
smithsonite
descloizite (cuprian)
dolomite
ferric hydrates and oxides
manganese hydrates and oxides

Galena, PbS. Isolated blebs of galena are found in the willemite rock towards the western end of the Zinc Reef. They are partially altered to black cerussite. Though generally larger than those of the "cluster zone" they are best interpreted as sulphide blebs originally disseminated in country rock.

Sphalerite, ZnS. Similar blebs of yellow-green sphalerite are common in the Zinc Reef proper; they are traversed by veinlets of the secondary minerals willemite, greenockite and unidentified black material. In places sphalerite has been totally removed as testified by cleavage boxworks of willemite.

Willemite, Zn_2SiO_4. This mineral is the dominant constituent of the Zinc Reef. Three modes of occurrence may be distinguished:

(*a*) *Compact crystalline willemite* forms the bulk of the rock. It consists of closely-packed hexagonal prisms with a granular or radiating texture. The aggregate may be greyish in colour with yellow, red, brown or black streaks of ferruginous material included in the crystalline willemite. More often the whole mass is permeated by red iron oxides occurring as inclusions and along grain boundaries of willemite.

It is believed that this type of willemite was formed by metasomatic replacement of dolomite and limestone. Occasionally a series of parallel sheet-like cavities in grey willemite are wholly or nearly filled with colourless willemite which has grown in compact radial aggregates from opposite walls. It is possible that these regular veins were filled bedding plane fractures in platy limestone (widened by shrinkage during replacement), thus supporting the replacement theory.

(*b*) *Drusy willemite* lines the walls of innumerable small veins and cavities in the compact willemite. Drusy willemite is also found along small fractures in the pink dolomite adjacent to the Zinc Reef.

The willemite crystals are slender needle-like prisms terminated at one end by a flat rhombohedron. They are colourless, or may be white in bulk. The crystallography, cleavage and other properties of these crystals are described below in some detail. The prisms range from about 0·5 to 3·0 mm. in length and 0·02 to 0·3 mm. in diameter, the average size being approximately 1·0 mm. × 0·12 mm. They occur singly, in diverging groups and in compact aggregates. Several modes of aggregation are recognisable:

Radial clusters are the commonest type. The needle-like crystals are usually in close contact and may form continuous layers with a botryoidal outer surface.

Parallel groups of short prisms form stout hexagonal clumps, sometimes distributed in radial form. These bunches are more or less isolated and measure about 1·0 mm. in length and 0·5 to 1·0 mm. in cross-section. The prisms are sometimes so closely grouped that the bunches assume the crystal form of single crystals with uneven prism and rhombohedral faces.

Carpet-like layers of flattened tabular crystals about 0·1 × 0·04 mm. in size form a cover on earlier willemite in open cavities.

Stalactitic growths with a radial structure are sometimes seen.

The spaces between fractured pieces of red willemite rock may be locally occupied by an oölitic aggregate of colourless willemite spheroids in close contact with each other. In thin section the rosette-like aggregates exhibit the extinction crosses characteristic of radial crystallisation.

(*c*) *Willemite boxworks:* Sphalerite blebs in the Zinc Reef are partially replaced along cleavage directions by black contaminated willemite. In areas where no sphalerite is preserved, black willemite stands out as linear or criss-cross ridges enclosing angular cavities. The ridges are often encrusted by minute drusy willemite crystals. Occasionally the tetrahedral cleavage pattern of sphalerite is recognisable in these boxwork structures.

Black willemite is only found where sphalerite has been replaced. Its colour is due to inclusions of the unidentified black mineral (possibly galena, see p. 275).

Chemical and optical properties: The clear, colourless, transparent nature of the crystals indicate that the willemite is pure. This was confirmed by a qualitative spectrographic analysis of a few selected crystals. No Mn was detected; boron is present as a trace element and was only found in one other mineral from Abenab West, viz. wulfenite.

The refractive indices of the willemite are as follows, as determined by immersion methods:

$$\omega = 1{\cdot}692 \pm {\cdot}0025$$
$$\epsilon = 1{\cdot}719 \pm {\cdot}0025$$

Abenab West willemite shows no fluorescence in the ultra-violet light of a mercury vapour lamp. It was also tested with the iron spark as a light source, but without success. Similar observations are recorded by L. J. Spencer (1927) for supergene willemite from a number of localities.

Crystallography: The morphology of the Abenab West willemite is simple and extraordinarily uniform. Only two forms were observed, namely a hexagonal prism terminated at one end by a flat rhombohedron of the same order. The rhombohedral faces were too small for accurate measurements with the reflection goniometer. However, an approximate value for the normal angle between prism and rhombohedron was found by applying two methods:

(*a*) A crystal was mounted with a drop of Canada Balsam on one of its prism edges on an object glass which was attached to the U.M. stage. The U.M. stage was used to set the zone axis of the faces to be measured parallel to the microscope axis and the desired angle, measured against an ocular cross-wire, was read on the microscope stage to the nearest tenth of a degree. Values so obtained from 12 crystals ranged from 63° 30′ to 67° 30′ and the average was 65° 15′.

(*b*) The Meirs reflection goniometer was combined with a binocular microscope and a crystal was attached to a needle point perpendicular to its c-axis.* With one nearly perfect crystal measuring $2{\cdot}4 \times 0{\cdot}35$ mm. the reflection positions of a collimated light source on the crystal faces could be determined to an accuracy of less than one degree. Values so obtained for 16 repeated measurements of all three prism/rhombohedron angles on one crystal ranged from 63° 53′ to 66° 20′ and the average was 65° 17′.

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* It has recently come to the author's attention that a similar but more precise procedure for measuring facial angles of minute crystals on a single-circle goniometer with the aid of a binocular microscope and an auxiliary light source has been described by T. Watanabe (Jour. Fac. Sci., Hokkaido Imp. Univ., Ser. IV, Vol. III, No. 2, 1936).

Measurements were rendered difficult by the fact that the small rhombohedral faces are usually uneven or slightly curved, though prism faces may be perfectly smooth. On the larger crystals the prism faces are usually striated vertically.

The normal angles lying in the vicinity of 65° between various prisms and rhombohedra of the same order were calculated from the axial ratio of willemite, given as 0·6775 by Dana-Ford, and the following values were obtained:

$$(10\bar{1}0) \wedge (20\bar{2}3) = 62^\circ\ 27'$$
$$(11\bar{2}0) \wedge (11\bar{2}3) = 65^\circ\ 42'$$
$$(10\bar{1}0) \wedge (10\bar{1}2) = 68^\circ\ 38'$$

Although the average angles measured are slightly below the theoretical value, it is obvious that the best correspondence is obtained if the prism is taken to be the second order prism $\{11\bar{2}0\}$ and the rhombohedron $\{11\bar{2}3\}$. This result is in agreement with the observation of an indistinct cleavage parallel to the prism faces.

According to Goldschmidt (1923), $\{11\bar{2}3\}$ is the only second order rhombohedron recorded for willemite. Both the negative form U$\{\bar{1}\bar{1}23\}$ and the positive form S$\{11\bar{2}3\}$ are present on a rather complicated crystal of equidimensional habit from Musartut, Greenland, the negative form attaining the larger dimensions. This rhombohedron is apparently not known from other localities, while at Abenab West it seems to be the only one to occur, either the negative or the positive form being present according to the orientation favoured.

Willemite crystals in every respect similar to those described above (except perhaps in abundance) have been described by L. J. Spencer from Broken Hill, Northern Rhodesia, whereas larger ones with the same crystal faces are reported from Sable Antelope Mine, Northern Rhodesia and Guchab, near Otavi, S.W.A. (Spencer, 1927). The crystals are also prismatic with rhombohedral terminations, the forms being the first order prism $\{10\bar{1}0\}$ and the negative first order rhombohedron e$\{1012\}$ according to Spencer. $\{\bar{1}012\}$ is one of the commonest forms of willemite and its theoretical angle with the prism is only 2° 56′ different from the angle between $\{11\bar{2}0\}$ and $\{11\bar{2}3\}$. Unfortunately no material from these localities was available in order to ascertain whether the forms present are definitely those of the first order, as reported by Spencer, or those of the second order believed to characterise the willemite from Abenab West.

The situation has been further complicated by F. H. Pough (1941) who substituted the second order prism $\{11\bar{2}0\}$ for Spencer's first order prism in listing willemite from Spencer's abovementioned localities, while retaining the first order rhombohedron $\{10\bar{1}2\}$. This would certainly be wrong for Abenab West, because at this locality the prism and rhombohedron are definitely of the same order. According to Pough, Lacroix described supergene willemite with the combination $\{11\bar{2}0\} + \{10\bar{1}2\}$ from four localities, while Pough himself adds three more with the "ubiquitous forms" $\{11\bar{2}0\}$, $\{0001\}$ and $\{10\bar{1}2\}$ as being the only ones present. Like the Abenab West mineral, these willemite crystals are small, ranging up to 2 mm. in length. No goniometric data on which the indices are based are given. Not one specimen with a single-order prism/rhombohedron combination is listed by Pough. Morphologically, the Abenab West willemite would therefore appear to be unique. The author suggests that this is not the case but that the commonest forms are in reality $\{11\bar{2}0\}$ and $\{11\bar{2}3\}$.

Cleavage: For two reasons special attention was focussed on the possible directions of cleavage in Abenab West willemite; firstly because previous observers have recorded marked differences between the cleavage of manganiferous willemite (Troostite) from New Jersey and ordinary willemite from Moresnet (Belgium), New Mexico and elsewhere; secondly because the presence of a prismatic cleavage would afford evidence for the determination of the order of the prism faces. The fracture of several single perfect crystals was carefully observed through the binocular microscope while they were being crushed between two glass plates. This procedure showed that fracture invariably took place first at intervals parallel to the basal pinacoid (which is not present), indicating that a {0001} parting is present. In the second place there is a definite tendency to break longitudinally for short distances along planes more or less parallel to the prism faces. This direction would seem to represent an indistinct prismatic cleavage. Sometimes fracture also takes place along planes steeply inclined to the c-axis.

The cleavage directions of Abenab West willemite are compared below with those given by Hintze for the two varieties mentioned above.

Ordinary	*Manganiferous*	*Abenab West*
Distinct {0001} clv.	Very indistinct {0001} clv.	Distinct {0001} parting (or cleavage?).
—	Distinct $\{11\bar{2}0\}$ clv.	Indistinct $\{11\bar{2}0\}$ clv.
—	Indistinct $\{10\bar{1}1\}$ clv.	—

As the only prismatic cleavage recorded is the one parallel to the second order prism, it seems likely that the prism faces of the Abenab West mineral are those of the second order. This conclusion confirms the results obtained by goniometric measurements. It must be noted, however, that the A.W. willemite resembles ordinary willemite in all respects except the presence of a prismatic cleavage, in colour, composition, habit, lack of fluorescence and mineral associations. Although Hintze regards the differences in cleavage and crystal form as characteristic enough to separate ordinary and manganiferous willemite into two mineral species, he gives the axial ratios as identical to the fifth decimal place. X-ray powder photographs (taken at the mineralogical laboratory of Prof. de Wijs in Delft, Holland) prove the identity of willemite from Abenab West and from Franklin Furnace, N.J. The unit cell may therefore be taken to be identical in both varieties, which makes it likely that the same cleavage directions will be present; their relative distinction depending on the paragenesis or crystal habit.

Zoning: When viewed in oblique light especially with the binocular microscope, most willemite crystals reveal dark or luminous lines which are interpreted to be the traces of growth zones analogous to those found in the small yellowish vanadinite crystals described earlier (Fig. 6). Near the rhombohedral terminations of the crystals these traces are parallel to the rhombohedral faces; lower down they are usually straight and parallel to the base. They are very closely spaced, but usually fade out and become prominent at intervals. Fracture under pressure was seen to proceed independent of the zonal traces, even when a {0001} trace was strongly developed. Nevertheless, crystals were seen with fractures parallel to the terminal rhombohedra, and these are ascribed rather to the presence of zoning than to a $\{11\bar{2}3\}$ cleavage. There is no zonal colour variation. It would seem that growth zoning is no less frequent in

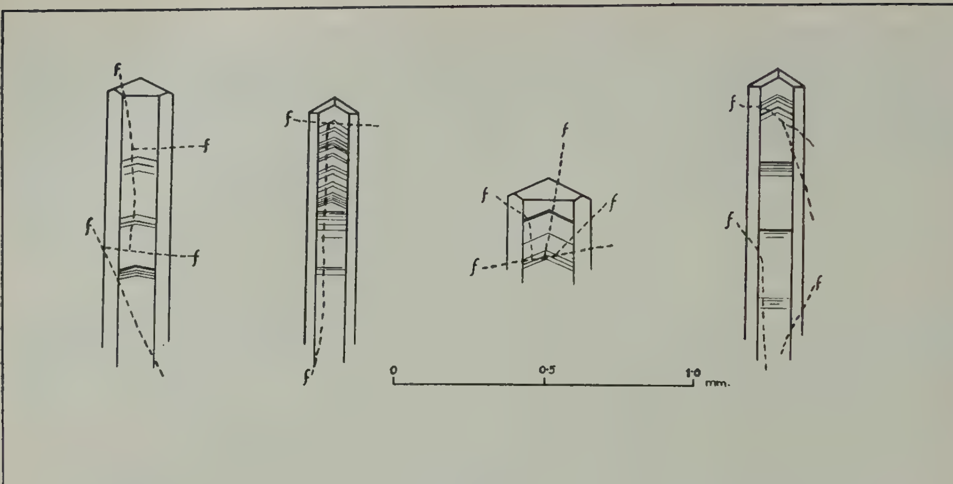

FIG. 6. Fracture and zoning of Willemite.

minerals with a fixed composition than in minerals with a variable composition such as descloizite.

Vanadian wulfenite, $Pb(Mo,V)O_4$

This mineral is a very rare constituent of the Zinc Reef. A few isolated grains, up to 2 mm. in diameter, were embedded in white drusy willemite. The yellow mineral was at first mistaken for vanadinite, but spectrographic tests revealed the presence of both Mo and V. An X-ray powder photograph was made, proving its identity with wulfenite.

The habit of the crystals vary from euhedral (thin prismatic or tabular) to subhedral; straight crystal faces truncate willemite spheroids in places and are interrupted by them in others. This indicates simultaneous crystallisation of wulfenite and willemite.

Under the microscope the grains are seen to possess a resinous luster and a sub-conchoidal fracture with one imperfect cleavage plane on which an eccentric uniaxial negative interference figure was obtained. Colour variations and marked zoning are apparent, there being an uninterrupted range between colourless non-pleochroic varieties through light yellow weakly pleochroic to orange yellow strongly pleochroic types. Regular streaks of the strongly pleochroic varieties occur in the more abundant lighter coloured types.

The orange to red colour of some wulfenite has been the subject of several investigations (cf. Hintze, 1930). P. Smith (Amer. J. Sc. 20, 245, 1853) was apparently the first to regard the presence of vanadium as responsible for the red colour of wulfenite from Phoenixville, Pennsylvania. Vanadian wulfenite has been named eosite. (Palache, Berman and Frondel, 1951.) In order to prove the presence or absence of vanadium in the wulfenite from Abenab West, all the available material was carefully crushed and the fragments separated under the binocular microscope into dominantly light yellow and dominantly orange-red varieties. On the 3 milligram and 7 milligram samples so obtained, Mr. W. J. Pienaar of the Western Province Fruit Research Station attempted the quantitative spectrographic determination of vanadium. The

results showed less than $0{\cdot}2\%$ vanadium in both samples, but cannot be regarded as reliable owing to the small quantity of wulfenite used. Abenab West wulfenite most probably owes its colour to somewhat higher percentages of vanadium.

Under high magnification numerous rounded and irregular inclusions (probably fluid cavities) were seen in wulfenite. The inclusions are grouped in streaks or lie along planes that cut across zonal boundaries. Rod-like inclusions about 0·01 mm. in length were also observed, sometimes in parallel orientation. Grains containing these inclusions were tested with the spectrograph, but showed no increased vanadium content. They are probably elongated cavities like the rounded and irregular ones with which they seem to be associated.

Greenockite, CdS

Citron-yellow grains of greenockite about 0·02 mm. in size are found to replace sphalerite blebs in the Zinc Reef along cleavage directions and superficial cracks. When associated with the ordinary secondary minerals like willemite, greenockite invariably intervenes between the sphalerite and its oxidation products. Stringers of greenockite occur parallel to veinlets of the unidentified mineral B (supergene galena?) but always on the sphalerite side.

Malachite, $Cu_2(OH)_2CO_3$

Grass-green prismatic crystals of malachite, arranged in diverging tufts, are confined within the walls of the angular cavities in some willemite box-works. Under the microscope they show pleochroism, one perfect cleavage direction with straight extinction, positive elongation and a lesser cleavage direction perpendicular to the first. Malachite is restricted to the sites of former sulphide ore and remnants of a black ore mineral (chalcocite?) were observed in one area. Malachite, in turn, is enclosed by later cerussite.

Cerussite, $PbCO_3$

The black variety surrounding remnants of galena is older than drusy willemite as it is encrusted by the latter. Colourless transparent cerussite with its characteristic oily lustre is sometimes found in drusy willemite and appears to be invariably younger. It may curve around willemite aggregates and fill a vug either partially or completely. Free surfaces of cerussite develop crystal faces. Euhedral crystals of cerussite, 1-3 mm. in diameter, were found supported by willemite crystals in the centre of a drusy cavity. Cerussite was never seen against the walls of a cavity also occupied by willemite. In one instance cerussite was found well within the massive willemite, occurring interstitially with respect to willemite clusters and red iron oxide pits.

Smithsonite $ZnCO_3$, is virtually absent from the Zinc Reef. A few very small rhombohedra were encountered in a cavity between the willemite ridges of a cleavage boxwork.

Descloizite, $Pb(ZnCu)OHVO_4$

Descloizite is one of the very last minerals deposited in the Zinc Reef. Only calcite has been found encrusting descloizite. It is not a common constituent and is very finely crystalline. Thin layers of tabular crystals, less than 0·1 mm. in diameter, encrust compact willemite, drusy willemite, dolomite crystals, calcite, malachite and fractured country rock. The descloizite varies in colour from light lemon-yellow to green, often with gradational boundaries between varieties. Clusters of larger olive-

green crystals on willemite were tested microchemically and showed approximately equal amounts of copper and zinc. Pseudomorphs of cuprian descloizite after vanadinite were found in the immediate vicinity of the Zinc Reef.

Calcite, $CaCO_3$

Transparent and white crystalline calcite forms the binding material of fractured willemite rock and fills many cavities lined by willemite and botryoidal oxides. Coarse zonal crystals are sometimes found. Layers of finely crystalline calcite have been deposited on drusy willemite, descloizite and other minerals in the Zinc Reef. The calcite invariably contains zinc and usually lead as shown by microchemical tests. (For analysis see Table 4.)

Dolomite, $CaMg(CO_3)_2$

A few small crystals of dolomite, characterised by high pearly lustre and yielding thin cleavage plates were found on botryoidal oxides. They are encrusted by descloizite and calcite.

Ferric hydrates or oxides

The willemite rock owes its red colour to a homogeneous dissemination of very fine ferruginous material. The red colouration can hardly be ascribed to subsequent infiltration but was deposited slightly in advance and incorporated by willemite. Grey compact willemite contains numerous red globules that weather out to form circular pits. In some cases they seem to represent centres around which willemite aggregates crystallised. In thin sections again, they are sometimes seen to spread over variously orientated willemite crystals and are therefore younger, at least in part. Botryoidal crusts of red iron hydrates with a colloform texture are found to be both older and younger than willemite. There are also thin layers deposited on drusy willemite and recent stains caused by superficial weathering.

Manganese hydrates or oxides

Black manganiferous layers, globules and botryoidal crusts are distributed in the Zinc Reef in exactly the same way as the ferruginous ones. Coatings of the manganese mineral (pyrolusite?) are especially abundant on the rhombohedral faces of willemite crystals. Manganiferous dendrites are quite common in the laminated country rocks shale, limestone and dolomite.

VI SULPHIDE MINERALISATION ALONG THE ABENAB WEST ZONE OF DEFORMATION

(1) *The sulphide ore of Drill Core SDDH*23

Sulphide ore, represented by a core length of 4 ft. 6 ins., was encountered in a surface diamond drill hole at a depth of 630 ft. below the surface. Although there is no proof of continuity with the oxidized "clay", this sulphide ore occurs at the same stratigraphic horizon and not improbably represents a deeper unoxidized portion of the present Abenab West ore body.

The core was split along its length and one half was divided in five portions, each of which was assayed for Pb, Zn, Cu and V_2O_5. The other half was available for mineragraphic study.

The main constituents along the whole length of the core are galena and sphalerite. Whereas galena seems to be unaffected yet by supergene oxidizing solutions,

sphalerite is partly replaced by secondary zinc minerals. Pyrite is rather uniformly distributed in both galena and sphalerite but the proportion between the latter two is highly variable. In the upper six inches of the core, some galena is associated with sphalerite which is riddled by drusy cavities and veins lined with smithsonite and pyrite. At 631 ft. there is an abrupt contact with massive relatively fresh sphalerite which continues down to 631⅔ ft. without appreciable galena. At this depth a rubble-filled cavity was apparently encountered, for all that were recovered from the next 1⅓ ft. were a few limonitic lumps studded with fragments of smithsonite, blende and galena. Below this cavity, relatively unaltered sphalerite and galena are intergrown in about equal proportions but with galena increasing towards the end. Four inches above the base of the sulphide ore, a half-inch vein of compact equigranular willemite cuts across sphalerite. The contact is marked by red streaks of iron oxide, next to which the sphalerite is extensively altered to smithsonite, the black unidentified mineral B and hydrozinkite. The relations between the secondary zinc minerals are not quite clear but it seems as if willemite is earlier than smithsonite.

The contents of the abovementioned cavity merit closer description because it yielded 3·30% V_2O_5 on analysis, in comparison with ·05-·10% in the rest of the core. Both upper and lower walls consist of a smithsonite boxwork associated with the greenish black copper arsenate and remnants of unaltered sphalerite, collectively encrusted by late, finely crystalline pyrite. On the pyrite lie loose fragments of galena, sphalerite and smithsonite (obviously not *in situ*) mixed with yellow limonitic material. Of the cavity filling only two pieces about ½″ in diameter were available:

(i) One is a soft orange-brown lump with a cavity lined by tiny reddish spheroids and twig-like structures of goethite. In the cavity, fragments have again accumulated, and among them several finely crystalline, lemon-yellow, rounded descloizite lumps were found.

(ii) The other piece is hard and compact, consisting of ore rubble cemented by zinciferous calcite contaminated by red iron oxide. Thin layers of finely-crystalline yellow descloizite traverse this material.

These observations tend to show that descloizite was deposited in the cavity before and during the period of rubble accumulation. The descloizite is not associated with primary sulphides but with secondary material as in the oxidized clay of Abenab West.

Although special attention was focussed on the possible identification of a primary vanadium mineral such as sulvanite or patronite, no such mineral was found in the polished sections that were studied.

The following minerals were identified:

Primary	*Secondary*
Sphalerite	Covellite
Galena	Greenockite
Pyrite	Smithsonite
Tennantite	Hydrozinkite(?)
Chalcopyrite	Willemite
	a copper arsenate

Sphalerite, ZnS. The heterogeneous colour variation of this finely-crystalline mineral is obvious even macroscopically. In thin section it is seen to consist of three components, viz. a colourless component in which irregular areas of a bronze and black variety, alternating in zonal fashion, are distributed. The variation is probably due to differences in iron content during crystallisation, but it is not quite clear whether

the colourless variety replaces the zonal type, or whether the irregularity is due to random orientation of interlocking crystals varying in colour. In vertically reflected light the sphalerite appears homogeneous. Structure etching with $KMnO_4 + H_2SO_4$ showed that it consists of a polygonal equigranular aggregate with abundant twin lamellae. The texture is similar to sphalerite from Broken Hill as illustrated by Ramdohr (1950, p. 347).

In contact with galena and tennantite, sphalerite grains and masses are typically rounded and idiomorphic, probably indicating an earlier start in crystallisation. The mineral is strongly triboluminescent (like sphalerite from Tsumeb) but does not fluoresce in the ultra violet light of a mercury vapour lamp.

The sphalerite is traversed by vuggy smithsonite. At the base of the sulphide mass it is extensively altered, ramified with cracks and stained black. Small unaltered grains of sphalerite are found in the willemite vein and are often enclosed in hexagonal willemite crystals.

Abenab West sphalerite, both from the drill core and from the disseminated blebs, differs from the Tsumeb mineral in cadmium content. Whereas the latter is characterised by exceptionally high percentages, e.g. 2·5% Cd (Schneiderhöhn, 1929), spectrograms of Abenab West material revealed only traces of the metal.

Galena, PbS, shows an interstitial relationship to sphalerite. Contacts are always concave towards galena. Where sphalerite is dominant, small anhedral areas of galena are confined within it. No secondary lead minerals were found replacing galena.

Pyrite, FeS_2, is distributed as isolated crystals in both sphalerite and galena, as minute specks in sphalerite and as larger masses and stringers traversing the ore. Pyrite has undergone hypogene replacement by sphalerite as shown by skeletal remains of cubic crystals, similar to the replacement relics in galena from Tsumeb. (Ramdohr, 1950, p. 575.) Elsewhere pyrite only occurs between adjacent grains of sphalerite. In the willemite vein and associated smithsonite and hydrozinkite, pyrite occurs as scattered crystals, partially or completely altered to red iron oxide and surrounded by red concentric rims at some distance from the original crystals.

Although most pyrite (occurring as described above) is definitely primary, a younger generation has been deposited on and along with smithsonite lining open cavities in sphalerite.

Tennantite, $(Cu, Fe)_{12}As_4S_{13}$. Patches of a greenish grey, isotropic mineral were observed in most polished sections. Its reflectivity, hardness and relief were found to be intermediate between those of sphalerite and galena. All etch reactions with the standard reagents recommended by Short (1940) were negative. Finally a spectrogram of material containing only a small proportion of the mineral proved the presence of Cu, As and the absence of Sb, V and Ag. It was concluded that the mineral is a relatively pure end-member of the tetrahedrite-tennantite series. The minerals of this series are often important for their silver content.

The tennantite occurs as disseminated grains in sphalerite or in galena at a sphalerite contact. It presents euhedral boundaries to galena but is concave towards sphalerite. Where it occurs in galena, it has the appearance of being moulded in sphalerite. Often tennantite occupies corners and embayments in sphalerite. Areas larger than 0·1 mm. in diameter are rare, but in places specks of about 0·02 mm. are abundant in sphalerite. A few droplets of sphalerite were observed as inclusions in tennantite. Tennantite was nowhere seen to have undergone alteration or replacement.

Chalcopyrite, $CuFeS_2$ occurs in galena as extremely small yellow inclusions measuring ·01-·02 mm. across. It has a high reflectivity, no relief and weak anisotropism. In the sections studied the chalcopyrite is very rare and only visible under high magnification and in oil. Nevertheless Cu and Fe were detected spectrographically.

Covellite, CuS occurs as secondary mineral along cracks in sphalerite and as blebs and veins in smithsonite and sphalerite.

Greenockite, CdS. A few citron-yellow grains of this mineral were seen, like covellite, in cracks in sphalerite.

Unidentified mineral B (supergene galena?). In thin sections of partly altered sphalerite, an opaque black mineral was usually seen to accompany the secondary zinc minerals. Minute, densely-crowded specks of this mineral are responsible for the black colour of smithsonite and willemite that penetrate the sphalerite along cleavage directions. The black mineral surrounds unreplaced remnants of the sulphide, sometimes separated by a thin band of intervening smithsonite. It also forms stringers along willemite veinlets like the marginal and medial moraines of a glacier. In shaded light it has a bluish-black metallic lustre. Sphalerite itself often shows an extensive black staining (not discernible in reflected light) which emanates from cracks and veinlets of the black mineral. The same black staining in sphalerite is also found along its contact with galena.

What is presumably the same mineral occurs in hand specimens of the sulphide core as bluish-black sooty layers along fracture planes. Mixed with sphalerite, this material gave a negative test for copper but a positive test for zinc and iron. A spectrographic determination indicated the presence of small amounts of lead, copper and iron in addition to zinc.

In polished sections the unknown mineral shows a galena-white reflectivity and is isotropic. Its identity with the mineral described above was proved by studying polished sections, thin sections and a polished thin section from the base of the sulphide core. Thin films of the mineral are moulded on sphalerite along its highly irregular contacts with smithsonite, while irregular specks are scattered within the adjacent smithsonite. It is clear that the mineral shows a typical supergene mode of occurrence and is closely related to the oxidation of sphalerite. The following etch reactions were obtained:

1 : 1	HNO_3	1 : 1 HCl	$FeCl_3$	positive
	KCN	$HgCl_2$		negative

Unfortunately the areas on a polished section are much too small to isolate for microchemical tests. Nevertheless, portions of a thin section containing the black material were tested for silver and tellurium but without positive results.

To the author the only logical conclusion seems to be that we are dealing with supergene galena deposited on sphalerite during a process of cementation.* Large anhedral patches of primary galena are present in the same polished sections and are indistinguishable from the thin fringes around sphalerite, even under high magnification and in oil. The blackening with HNO_3, the brown colouration with HCl and the iridescent staining with $FeCl_3$, are characteristic of both. According to Ramdohr (1950, P. 453), galena is only very rarely redeposited during weathering,

* This conclusion is confirmed by the fact that supergene galena has been identified in ore from Broken Hill, S. Rhodesia, as described and amply illustrated in a paper by J. H. Taylor which appeared two years after the present study was completed. ("Colonial Geology and Mineral Resources," Quart. Bull. Colonial Geol. Surveys, Vol. 4, No. 4, London, 1954). Taylor also reaches conclusions similar to those of the author concerning the origin of the vanadium.

and apparently only from highly acid solutions as in the case of galena with gel structure on sphalerite.

Smithsonite, $ZnCO_3$, is the principal secondary mineral in the drill core. Clear, transparent crystals line veins and cavities in sphalerite. In compact, partly oxidized ore, replacement remnants of sphalerite are surrounded by impure smithsonite crowded with inclusions of the accompanying unidentified black mineral. Interstitial patches of clear smithsonite are found between willemite crystals.

Hydrozinkite, $Zn_5(OH)_6(CO_3)_2$ (?). In thin sections of the partly altered sphalerite, extensive areas of a colourless fibrous mineral were observed in close association with smithsonite. The individual laths measure approximately $0 \cdot 12 \times 0 \cdot 02$ mm. and are grouped in massive radial aggregates, often with a core of red iron oxide. The aggregates form an interlocking mosaic pattern just like smithsonite, and often enclose patches of the latter. Possibly smithsonite is being replaced by the fibrous mineral, which is provisionally identified as hydrozinkite. It has the following properties:

Positive elongation and straight extinction of laths. Refractive indices:

$1 \cdot 68 \pm \cdot 01$ perpendicular to the elongation.

$1 \cdot 73 \pm \cdot 01$ parallel to the elongation.

The mineral dissolves slowly in cold 1 : 7 HNO_3, effervesces in hot acid and gives a strong zinc reaction. Tests for Cu, Pb, Ca, As were negative.

Willemite, Zn_2SiO_4, was only found in a vein, half an inch thick, near the base of the sulphide core. In thin section it forms an equigranular aggregate with a brownish tinge and ramified by imperfect cleavage cracks. Hexagonal basal sections are common and zonal cracks are sometimes seen in them. The willemite occurs as irregular areas and veins in sphalerite, without an obvious replacement relationship. Perfectly clear willemite is found in contact with black contaminated smithsonite that replaces the sphalerite. Willemite is always euhedral towards smithsonite and is thought to be earlier. It was probably deposited by siliceous solutions migrating from elsewhere.

A green *copper arsenate* (olivenite?) occurs in powdery form in an open cavity lined by smithsonite and associated pyrite. Optical properties could not be determined.

(2) *The Disseminated Sulphides in the "Cluster Dolomite".*

The pinkish laminated dolomite horizon which forms the hanging wall of the Abenab West ore body contains a central zone, about 60 ft. thick, known as the "Cluster Dolomite". It is characterised by numerous groups (1-2 cm. in diameter) of white dolomite crystals wholly or partly filling small cavities in the dolomite rock. The "cluster zone" maintains a greater regularity than some of the stratigraphical units and has been traced continuously for more than ten miles across the mining area.

A zone of disseminated sulphides coincides roughly with the Cluster Dolomite, although it usually transgresses across the lower pink dolomite also. The sulphide blebs do not form a continuous zone but tend to be concentrated in particular areas with narrow lenticular ground-plan. Nevertheless there is a close relationship between the dolomite clusters and sulphide blebs. They are usually comparable in size and very often the sulphide occurs interstitially in a dolomite cluster. Both zones are truncated by some of the arcuate faults of the Abenab West horizon and are therefore certainly older than the faulting.

According to visual estimates the disseminations average about 2% sphalerite and galena together. In drill cores it was found that where the disseminated mineralisation is most prominent the sphalerite and galena usually occur together and are some-

times intergrown with each other. In many drill holes the galena and blende are segregated from each other in narrow sub-zones separated by barren patches.

A detailed study of the distribution of the disseminated sulphides in the Abenab West mine revealed a remarkable correspondence with the ore body. Where the mineralised clay dies out at the eastern and western ends of the 200 ft. level, the disseminated sulphides also become scanty and seem to tail off in narrow lenses. The relative proportion of disseminated galena to sphalerite is greater in the central part of the mine and decreases towards both ends of the 200 ft. level, where only sphalerite was seen. Sphalerite predominates on the whole, but between W33 and E10 on the 200 ft. level galena is plentiful and sphalerite lacking. The mineralised clay is likewise characterised by a central part with a high lead content.

In addition to the Cluster Dolomite, the so-called Breccia-limestone horizon occasionally carries galena and sphalerite as stringers and specks in the groundmass.

The *mineralogy* of the disseminated sulphides is not very different from that of the drill core sulphides.

Galena: Galena has a normal granitoid texture. It is usually rimmed by cerussite, but may appear unoxidized. Blebs of galena as much as 10 cm. in diameter were seen on the 200 ft. level where the disseminated sulphide zone transgresses on to the Blue Dolomite.

Sphalerite: Two varieties of sphalerite in the sulphide blebs may be distinguished in the field: a yellowish green variety and a reddish brown one. The yellowish green sphalerite is the most abundant type, but in some areas the blebs consist solely of the brown variety. They are often found together, the brown variety occurring as zonal streaks in green granular sphalerite. In polished sections the two varieties may be distinguished to some extent by the colour of their internal reflections. Spectrographic determinations on both varieties revealed no manganese, a roughly equal iron content and appreciable lead in the yellowish green variety, possibly due to admixed galena. The colour differences therefore remain unexplained.

Sphalerite is usually interstitial to white dolomite crystals; in a thin section sphalerite was also seen to replace dolomite along its cleavage directions.

Sphalerite blebs may be perfectly unaltered, but are more usually rimmed and veined by willemite. Small cleavage boxworks of willemite, much contaminated by ferruginous material, often prove the former presence of sphalerite blebs in the dolomite. These boxworks are sometimes encrusted by yellowish rhombohedra of smithsonite, a subordinate alteration product of the sulphide. Greenockite and the unidentified mineral B are also found among the oxidized minerals of the sphalerite blebs.

Pyrite: Pyrite occurs as minute inclusions and as larger idiomorphic crystals in the sphalerite.

In addition to its association with sphalerite and galena blebs in the Cluster Dolomite, pyrite is widely disseminated in shale and limestone in the vicinity of Abenab. Single crystals 0·1-5·0 mm. in size are often scattered parallel to the bedding planes and may be either fresh or oxidized to limonite. On the 200 ft. level of the mine a bed of bluish grey laminated limestone was found to be densely crowded with minute pyrite crystals up to 0·3 mm. in diameter. In thin section each of these crystals was seen to have two fan-like borders of a colourless mineral on opposite sides. Each border consists of a bunch of parallel or contorted bands perpendicular to the pyrite crystal faces. The borders of all the pyrite grains in the section are orientated in one direction, and their outlines tend to imitate the crystal form of the pyrite cores. The colourless mineral has about the birefringence of quartz, $\pm$ elongation and undulose extinction.

VII OTHER MINERAL DEPOSITS AT ABENAB

(1) *The Karuchas sulphide zone*

This is a zone of silicification and associated sulphide mineralisation in the upper part of a grey massive dolomite horizon a few hundred feet north of the Abenab fault. A network of irregular quartz veinlets occupies cracks probably formed by crushing and brecciation along a specific lithologic horizon. The Karuchas zone extends over a distance of more than twenty miles. Galena and dark brown sphalerite occur at many places, usually in linear zones from a few to several hundred yards in length. The sulphides appear as disseminated crystals, in cracks and along larger dolomite crystals. In the larger zones polygonal aggregates of galena predominate over sphalerite and locally galena may be the only mineral as at Karuchas. According to an assay, galena from this locality contained 7·1 oz./ton silver. At the smaller occurrences blende predominates over galena.

Three-quarter mile west of Abenab on this horizon, relatively high grade disseminated galena occurs in narrow lenses only a few feet wide but persisting over several hundred yards in a strike direction. A polished section of galena from this locality revealed the presence of *bournonite* inclusions with polisynthetic twinning. The fresh galena receives a purplish blue tarnish on exposure.

(2) *The old Abenab breccia pipe and Vanadium mine*

This important deposit has been described by C. M. Schwellnus (1945) who regarded the breccia as a cave collapse phenomenon. The ore body took the form of a roughly cylindrical pipe of brecciated country rock steeply inclined in the dip direction of the enclosing platy limestone. It was situated along the overthrust limestone-dolomite contact which is also inclined to the NNW. The breccia pipe has been followed to a depth of 747 ft. and the bottom has not been reached. In detail the downward course of the body was step-like as shown in Fig. 2 (AA′) which illustrates the shape of the breccia pipe and its relation to the Abenab West mine.

Compact reddish clay and crystalline aggregates of descloizite intimately associated with calcite formed the binding material of the breccia blocks. Interstitial cavities up to several feet across were partially or wholly filled by large calcite, descloizite and occasional vanadinite crystal groups. In the platy limestone immediately adjoining the breccia body, small discontinuous stringers of descloizite filled cracks and joints. Below 700 ft. the breccia body rapidly narrowed downwards and the matrix became a red unconsolidated clay.* (Unfortunately the mine was closed before the author's visit to the area, so that isolated specimens from older collections had to be utilised for a brief mineralogical study.)

Descloizite: Descloizite is the only ore mineral in the majority of specimens from Abenab. Compact crusts of dark green descloizite coat individual breccia fragments and are usually succeeded by crystal aggregates from a few millimeters to several centimeters in length. These crystal aggregates are usually enveloped by later calcite but may occasionally project into open cavities. They consist of numerous pyramidal crystals in sub-parallel or branching intergrowth with uneven, dull, black faces. In smaller sheet-like cavities, the drusy descloizite is lighter green in colour and tabular

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* Vanadium mineralisation continued below this depth as proved by short diamond drill holes from the 11th level, but the concentration appeared to be less and it was decided at the time not to sink the shaft any further, owing mainly to the high cost of pumping. A recent diamond drill hole from the surface has proved the continuation of the Abenab mineralised breccia to a depth of at least 889 ft. This is by far the greatest depth at which vanadates have been found in the Otavi Mountain Land.

in form. Later crops of descloizite may encrust drusy calcite crystals or are enclosed in a groundmass of clear, transparent calcite.

In thin section, and even macroscopically, the descloizite shows striking colour variations. Along the margin of a cavity, where deposition started, the descloizite is usually dark green and compact, but towards the centre a single crystal aggregate may become successively greenish yellow, golden yellow, orange and cherry red. The strongly pleochroic red variety is often found as a mesh of pyramidal crystal aggregates up to a few centimetres in size, intimately intergrown with calcite. All Abenab descloizite is built of regular zones varying in colour and pleochroism. While an orange red component was dominant in certain places (notably towards the end of crystallisation), the most abundant variety is lemon-yellow under the microscope and forms the groundmass in which orange, opaque yellow and blackish zones are distributed parallel to the crystal faces. The exact nature of the zoning is more fully described elsewhere.

Vanadinite: According to A. W. Clark (1931) "a considerable quantity of vanadinite is found below the fourth level (at a depth of 120 ft.) in clusters of large prismatic crystals up to three or four inches long, usually coated with a film of descloizite as much as 1/16 inch thick."

This exceptionally coarse vanadinite is brilliant red in colour, giving light orange-coloured powders and thin sections. Under the microscope it shows a zonal alternation of yellow and orange red pleochroic layers, but with weaker pleochroism than the red descloizite. It can also be distinguished from the latter by its lower birefringence. The absorption is $\omega > \epsilon$, fracture is sub-conchoidal and lustre on broken surfaces is resinous and higher than descloizite. Uniaxial and slightly biaxial negative interference figures were obtained on crushed fragments. A chemical analysis (Table 2) indicated the presence of $2{\cdot}91\%$ As_2O_5 as isomorphous replacement of V_2O_5.

Some of the red vanadinite prisms are arranged in diverging groups and were probably found as cavity linings. Pyramidal faces are occasionally well-developed and the two forms $\{10\bar{1}1\}$ and $\{20\bar{2}1\}$ could be identified by measuring the interfacial angles with a contact goniometer. Green descloizite encrusts these crystals but was not found to replace them. Concentric cracks due to weathering are formed parallel to the zones in coarse vanadinite.

Calcite: Together with the vanadates, calcite forms the principal binding material of the breccia. It is transparent and may be coarsely crystalline, forming crystals up to three cms. in diameter in drusy cavities.

Siderite: Siderite, associated with white *dolomite* crystals, was found to fill interstices between the fragments of a blue dolomite breccia picked up on the old Abenab waste dump. It is doubtful whether this is derived from the old Abenab mine.

(3) *The Okarundu breccia pipes*

One and a half miles west of Abenab two small breccia pipes of the Abenab type are found. The larger of the two has a roughly circular ground-plan with a diameter of about 150 ft. and is situated on the Abenab fault horizon marked by a shear breccia. Fragments of the fault breccia lie in the pipe breccia which mainly consists of closely packed angular fragments of limestone cemented by coarse transparent calcite. A 50 ft. shaft and several drill holes revealed only traces of vanadium mineralisation.

Another clacite-bearing pipe-like breccia occurs about 100 yards south at the basal contact of the limestone. It is oval in plan and shows no shearing effects. The breccia carries appreciable descloizite deposited in alternation with calcite between the limestone fragments. It has been proved to a depth of about 250 ft.

Descloizite is the only ore mineral found in this breccia body. It was invariably the first to be deposited on the limestone fragments and is separated from a second crop of descloizite by clear transparent calcite. The descloizite is dark green in colour and apparently homogeneous. An analysis (Table 3) shows it to be moderately cupiferous and not very different from Abenab West descloizite. Microscopic investigation showed that this descloizite like the rest is not homogeneous at all but consists of a lemon-yellow component containing thin orange zones and black wedge-like streaks, while in most crystals a fourth, colourless to grey component occupies a peripheral position.

(4) *The Okarundu mineralised Breccia-limestone*

The so-called "breccia-limestone" is a thin persistent layer between the Blue Dolomite and the footwall shale band. It is 1-4 ft. thick and most probably of a sedimentary origin. It consists mainly of limestone fragments up to 1 cm. in diameter cemented by calcite, compact reddish clay and a little quartz. The fragments are closely packed, extremely irregular in shape, and are quite unsorted and unbedded. This horizon must have been particularly susceptible to mineralisation, for in the Abenab West mine the groundmass sometimes contains galena while at Okarundu a small vanadium deposit is localised within the breccia-limestone.

This occurrence is close to the small vanadium-bearing breccia pipe in the Far West Hills $1\frac{1}{2}$ miles from Abenab. It appears to persist to a shallow depth only. The mineralogy of this locality is of interest because the relationship between coarsely crystalline vanadinite and descloizite is so clearly shown.

Vanadinite must at one time have been the dominant or perhaps the only vanadium mineral at this locality. It is found as yellow to orange remnants of stout hexagonal prisms ·5-1 cm. in length, enclosed by pseudomorphic shells of finely crystalline descloizite. The vanadinite is partly decolourised and disintegrated, often consisting only of fragile concentric shells reflecting the zonal structure of the fresh mineral. Veinlets of descloizite were not seen to penetrate into the vanadinite.

Descloizite occurs as minute greenish to brownish crystals forming thin crusts on limestone fragments and vanadinite. Where the latter has been leached the crusts of descloizite form a hexagonal mesh enclosing cavities filled by calcite. Isolated shells of descloizite with the form of a hexagonal prism are also found in calcite and consolidated clay. Most , if not all, of the descloizite at this locality appears to have crystallised at the expense of vanadinite; but no complete and solid pseudomorphs were found.

Microscopically the descloizite was seen to contain a high proportion of the orange red pleochroic component as zones in a yellow-green matrix. In a few specimens the descloizite crusts were of a bright grassgreen colour, quite different from the usual type, and proved to have an appreciable copper content, though not in excess of zinc. The bright green cuprian descloizite also encrusts acute rhombohedra of calcite in the groundmass.

Calcite cements the whole mass of limestone fragments, consolidated clay, vanadinite remnants and pseudomorphic descloizite crusts into a heavy compact rock.

VIII DISTRIBUTION OF ORE MINERALS IN THE ABENAB WEST MINE

The economically important ore minerals in the Abenab West mine are galena cerussite, vanadinite, descloizite and willemite. All of these attain their maximum development at specific localities in the ore body, although descloizite seems to be a ubiquitous constituent. One aim of the present investigation was therefore to deter-

mine the quantitative distribution of the ore minerals and to correlate the mineralogical composition of the ore body, if possible, with its chemical composition as revealed by assay plans. For this purpose a series of 68 representative samples was personally

TABLE 6: TRACE ELEMENTS IN MINERALS OF THE ABENAB AREA

Mineral.	No. of Spectrograms.	Little.	Very little.	Doubtful.
Descloizite...	27	Al, Sn, As	Ge, Be, Ni, Fe	Ga, Mn, Ti, Cd, Co, Ag
Vanadinite (Abenab) ...	3	Al, Sn, As	Ge, Be, Ni, Co	Ag
Vanadinite (Abenab West)	2	Al, Sn	As, Ge, Be, Ni, Cu	Ag
Wulfenite	2	V, Al, Sn, B	Ge, Fe, Ni, Cu	
Willemite	2	Pb, Fe, Mg	Al, Ge, Sn, B, Cd, Cu	
Sphalerite	5	Pb, Cu, Fe, Si, Cd, Mg	Al, Ge, Mn	Ag
Galena	2	Ga, Cu, Fe, Si, Ag, Mg	Al, Ge, Mn, Sn, Cd	

NOTE: Varying amounts of Si, Mg and Ca are also present in the secondary minerals, probably in fluid inclusions (See Chapter X).

collected by the author in all accessible parts of the mine. These were subjected to laboratory study and were assayed for the three principal chemical constituents lead, zinc and vanadium.

In addition to the abovementioned ore minerals small quantities of other primary constituents (pyrite, sphalerite, bournonite) and oxidation products (covellite, anglesite, mimetite) are present, while the gangue consists of hydrated ferric oxides, clay, sand, calcite and dolomite rock. The possible effects of these on assay values of mine samples had to be investigated.

Sampling: Channel samples were cut along the floor, walls or roof of cross-cuts traversing the mineralised clay perpendicular to its strike direction. Owing to the soft nature of the clay, no difficulty was encountered with sampling. Where exposed, the whole thickness of the ore body from hanging wall to footwall was included in one sample. No attempt was made to study mineralogical variations in transverse sections, as the mineralised clay is seldom more than a few feet thick and because consecutive assay values in this direction do not show significant variations.

The samples were dried in an oven when wet, crushed with a steel mortar and pestle and screened through a 10 mesh (I.M.M.) sieve. In order to keep minute disseminated crystals as far as possible intact, the fine material was screened off first and only the coarser lumps were crushed. Whole samples, treated in this manner, weighed from 5 lb. to 10 lb. and had to be reduced with a mechanical splitter. Three representative fractions, weighing about 1 lb. each, were obtained from each sample: one labelled (*b*) was intended for mineralogical study, another labelled (*d*) was left at Grootfontein for reference purposes, while the third was ground to about 200 mesh in the mill used in preparing mine samples for assaying. This finely ground portion was again split into two representative fractions: (*a*) which was passed on to the assay laboratory and (*c*) which was kept for possible complete analyses. Eventually it was found that the finely ground portion (*c*) had a very convenient grain size for micro-

TABLE 7 (*a*): EFFICIENCY OF SEPARATION BY PANNING

	C5 (*c*)			D10 (*c*)			F7 (*c*)			G10 (*c*)		
Ore Minerals in Concentrate.	Cerussite *** Willemite ** Vanadinite * Descloizite *			Vanadinite **** Descloizite **			Galena ** Cerussite **** Vanadinite ** Descloizite ***			Descloizite *		
Percentage weight of clay fraction (X)	37·3			41·0			35·3			48·2		
	Pb	V_2O_5	Zn	Pb	V_2O_5	Zn	Pb	V_2O_5	Zn	Pb	V_2O_5	Zn
Assay on original (*c*) sample (Y) ...	5·39	0·70	3·49	26·99	7·82	1·4*	28·12	0·98	1·51	2·79	0·91	2·20
Assay on clay fraction (Z)	3·85	0·92	4·10	17·35	5·43	0·81	9·64	0·88	2·20	3·07	1·15	2·47
Loss ... ... $\frac{(X \times Z)}{(100)}$	1·44	0·34	1·53	7·10	2·23	0·33	3·40	0·31	0·78	1·48	0·55	1·19
Percentage Loss $\frac{(\text{Loss} \times 100)}{Y}$	26·7	48·6	43·8	26·3	28·5	23·6	12·1	31·6	51·7	53·0	60·4	54·1

**** Dominant. *** Abundant. ** Common. * Present.

* 0·07% Zn as reported is clearly in error; the more likely value 1·4% Zn is taken from the assay of Sample D10 portion (*a*). (Other assay values above show good correspondence with (*a*) samples (compare Table 11).

TABLE 7 (*b*): RATIO BETWEEN ASSAY VALUES BEFORE AND AFTER PANNING

	Original Sample (*c*).			Clay Fraction.		
	$\frac{Pb \times 100}{Pb + V_2O_5 + Zn}$	$\frac{V_2O_5 \times 100}{Pb + V_2O_5 + Zn}$	$\frac{Zn \times 100}{Pb + V_2O_5 + Zn}$	$\frac{Pb \times 100}{Pb + V_2O_5 + Zn}$	$\frac{V_2O_5 \times 100}{Pb + V_2O_5 + Zn}$	$\frac{Zn \times 100}{Pb + V_2O_5 + Zn}$
C5 (*c*)	56·3	7·3	36·4	43·4	10·4	46·2
D10 (*c*)	74·5	21·6	3·9	73·5	23·1	3·4
F7 (*c*)	91·9	3·2	4·9	75·8	6·9	17·3
G10 (*c*)	47·3	15·4	37·3	45·9	17·2	36·9

scopic work, and nearly all the mineralogical determinations were accordingly made on this portion instead of on (*b*) which was too coarse for this purpose.

Experimental Procedure

A preliminary investigation of certain experimental gravity table concentrates supplied by the Reduction Officer at Abenab, indicated that a fairly accurate estimate of their mineralogical composition could be arrived at by grain counting after the constituent minerals had been correctly identified. In the quantitative study of the mine samples, however, a serious problem had to be solved before grain counting could be attempted. In every sample, the mineral grains of ore and gangue are mixed with a large amount of amorphous ferruginous clay. The difficulty is to obtain a clean concentrate fit for the microscope, without appreciable or differential loss of ore minerals.

The following methods of separation were tried out in order to establish a suitable routine treatment of the representative samples prior to the quantitative study:

(1) Screening through very fine sieves (e.g. 325 mesh).
(2) Panning and decantation with water.
(3) Separation by specific gravity using Clerici's solution.
(4) Magnetic separation with the Hallimond Electromagnetic Separator.

The only method that succeeded in giving a clean concentrate was panning and decantation. The others failed to achieve a satisfactory separation of amorphous material, especially when the latter is attached to the mineral grains. However, panning is only a rough method of separation, not recommended for quantitative results (Twenhofel and Tyler, 1941). It was found that minute crystals and fragments of vanadinite and descloizite were continually lost in the process, no matter how carefully it is carried out. The loss was determined by analysis in the case of a few typical samples differing in mineral composition. (Table 7 (*a*) and (*b*)).

From Table 7 it is clear that more than 50% may be lost in samples with a low grade of ore, while the average loss is in the vicinity of 30%. In samples containing relatively coarse cerussite and galena together with relatively fine vanadinite and descloizite, there is a differential loss of vanadium minerals. In samples containing vanadium minerals only, the $Pb/V_2O_5/Zn$ ratio is more or less the same in clay fractions and original samples, though a slightly greater loss of V_2O_5 is apparently indicated.

While panning therefore succeeds in giving a clean concentrate, the method is not suitable for quantitative work. Two more attempts were made to achieve a quantitative separation of mineral grains.

(5) *Chemical Methods:* The problem under consideration would be solved if conditions could be found under which hydrated ferric oxides are dissolved whereas vanadinite, descloizite, cerussite and galena remain unaffected. Truog, Taylor and others (1936) described a chemical method for the removal of free iron oxide, free alumina and colloidal silica from soils without seriously attacking minerals like apatite. Sodium sulphide and oxalic acid are used to produce nascent hydrogen sulphide which changes the iron oxide by surface action at 80°-90°C. to sulphides of iron. These dissolve readily in dilute oxalic acid or HCl with an acidity as low as pH 3·5.

The detailed procedure of Truog, Taylor and others was followed with a typical sample of Abenab West ore. Unfortunately it was found that (*a*) vanadinite and especially cerussite were immediately covered by black PbS, making them scarcely distinguishable from other opaque black material in the ore; (*b*) oxalic acid alone tends to deposit small white crystals (probably lead oxalate) on cerussite and vanadinite,

rendering them opaque; (*c*) the black sulphidized "clay" was not sufficiently dissolved even with 0·05 NHCl, when the ore minerals are already appreciably attacked. Experiments with H_2S, NH_4OH, dilute HCl and salicylic acid met with no greater success. Attempts were also made to facilitate solution with the aid of electrolysis.

No satisfactory chemical method for the removal of the ferruginous clay could be developed as a result of these experiments.

(6) *Flotation:* As a last resource, five mine samples differing in mineral composition were submitted to the Reduction Officer at Abenab with the request that the quantitative separation of the ore minerals should be attempted by flotation on a laboratory scale. The flotation tests proved not too successful as was shown by analysis. The Reduction Officer stated that "the flotation of finely ground and clayey samples, even if entirely successful in separating the bulk of the wanted minerals, would not produce clay-free concentrates."

The flotation method failed to yield clean concentrates while the recovery was not sufficiently high for the quantitative purposes in mind.

Conclusion: Of the various possibilities investigated, no method of separation proved to be quantitative enough for grain counting on representative concentrates. One possible solution to the problem, namely the use of a Super-panner, could not be tested because the apparatus was not available. In view of these results it was decided that the idea of a quantitative survey of the ore body, involving the percentage determination of each ore mineral in the various parts of the mine, would have to be abandoned. However, it would be possible to estimate the relative proportions of the ore minerals to an approximate degree in the clean concentrates obtained by panning. Once the qualitative mineralogical composition of an ore sample is known, it might be possible to calculate mineral percentages from the assay values or from a full analysis. Accordingly the procedure adopted in the systematic study was as follows.

The representative portions of the mine samples, labelled (*c*), were reduced by coning and quartering to a weight of 20-30 grams. Each fraction was thoroughly wetted in a white porcelain evaporating dish, agitated in water and rubbed with the fingers in order to liberate the mineral grains as far as possible from adhering clay. After settling of the heavy grains, the mud and water were carefully decanted. This procedure was repeated until the supernatant liquid remained clear. Care was taken to lose as little as possible of the mineral grains. By giving the dish a rotary motion, the heavy ore particles were concentrated in the centre of the dish while the gangue fragments collected on top. The clear concentrate was dried on a water bath and eventually studied under the microscope by immersion in convenient liquids. A number of slides were taken from different portions of the concentrate before an estimate of the mineralogical composition was made. The ore minerals were classified in four groups under the headings: dominant, abundant, common and present. Grains of ore minerals other than galena, cerussite, vanadinite, descloizite and willemite were not encountered in the representative samples.

Results of Systematic Study.

The ore mineral content of 68 representative samples, as determined by estimation on panning concentrates, is listed in Table 8. These estimates have little quantitative value owing to the rough method of concentration adopted, but they may be used to construct a picture of the qualitative distribution of the ore minerals in the Abenab West mine. By plotting the results on a longitudinal projection of the ore body, a marked zoning of the ore minerals is made apparent. As a result of the

TABLE 8: COMPOSITION OF REPRESENTATIVE SAMPLES FROM ABENAB WEST MINE

Key: **** Dominant ** Common
*** Abundant * Present

Sample	Location	% Pb	% V_2O_5	% Zn	Ore Minerals in Concentrate				
					Galena	Cerussite	Vanadinite	Descloizite	Willemite
	50 ft. level								
C1	W15	26·4	7·0	1·5			****	**	
C2	W17	42·6	5·4	1·1		****	***		
C3	E02	9·3	0·4	1·5	*	**	*	*	
C4	E06	23·3	0·3	1·0		****		*	
C5	E29	5·7	0·8	2·9		***	*	*	**
C6	E32	2·3	0·2	24·7		*			****
C7	E15	7·4	2·9	4·5			**	***	**
C8	E12	8·4	1·6	3·4		**	**	**	
C9	E09	11·6	0·6	1·8	*	***		**	
C11	W38	28·4	3·3	<0·1	***	****	**	*	
C12	W34	24·0	2·9	1·0	***	****	**	**	
	100 ft. level								
D1	W03	52·0	1·14	0·8		****	*	*	
D2	E06	3·2	1·2	6·0		*	*	**	
	75 ft. level								
D3	E08	29·3	5·0	1·0		****	*	*	
D4	E07	19·7	8·3	6·6			*	****	
	86 ft. level								
D5	E14	18·3	1·5	1·5		****	*	**	
	90 ft. level								
D6	E19	6·7	3·0	3·4				**	
D7	E23	1·8	0·77	1·5				*	
D8	E29	10·1	4·34	4·4			*	***	
D9	E33	4·4	1·61	2·8				**	
D10	W19	27·0	7·56	1·4			****	**	
D11	W20	40·3	8·34	1·9		***	****	**	
D12	W24	18·6	4·91	4·7			***	*	
D13	W39	5·3	2·12	1·3			*	**	
D14	W46	2·6	1·16	1·0			*	*	
D15	W55	8·4	3·71	3·5				***	
	150 ft. level								
E1	W15	23·2	7·25	3·3			****	****	
E2	W20	36·2	3·78	1·1	*	****	*	*	
E3	W02	17·8	0·74	1·3	*	****		**	
E4	E09	9·1	3·28	6·9				****	
E5	E08	5·8	1·11	3·0		***		**	
E6	E03	19·1	0·58	3·7		****		*	**
E8	W25	31·9	3·10	1·1	**	****		*	
	200 ft. level								
F1	E21	12·5	4·99	3·7				****	
F2	E07	5·7	2·39	2·4				***	
F3	W07	8·2	2·00	3·3		**	**	***	
F4	W20	44·5	0·55	1·5	**	****		*	
F5	W23	22·1	1·72	2·4		****	**	**	
F6	W25	24·9	1·55	2·0	*	****	*	*	
F7	W28	29·2	1·18	2·0	**	****	**	***	

TABLE 8: (*Continued*) COMPOSITION OF REPRESENTATIVE SAMPLES FROM ABENAB WEST MINE

Key: **** Dominant ** Common
*** Abundant * Present

Sample	Location	% Pb	% V_2O_5	% Zn	Ore Minerals in Concentrate				
					Galena	Cerussite	Vanadinite	Descloizite	Willemite
	186 ft. level								
F8	W33	32·9	2·87	1·6		****	**	**	
	170 ft. level								
F10	W27	39·9	16·02	10·6				****	
F11	W30	21·2	5·53	2·0			****	**	
F12	W33	27·2	1·61	1·5		****	**	**	
F13	W35	29·0	12·73	7·9			*	****	
F14	W47	3·7	1·35	1·7		*	*	**	
F15	W53	19·1	8·17	5·9			*	****	
F16	W57	19·0	5·81	2·7			****	***	
F17	W68	1·0	0·47	1·6				**	
F18	W26	26·4	5·20	3·1		***	***	***	
	340 ft. level								
G1	W35	8·3	3·65	2·0				***	
G2	W39	11·1	4·79	3·4				****	
G3	W42	6·9	2·97	2·2				***	
G4	W45	3·1	1·40	0·6				**	
G5	W52	3·8	1·90	1·3				**	
G6	W30	5·5	2·71	2·1				**	
G7	W27	23·6	10·24	6·3			*	****	
G8	W14	16·3	7·12	4·9				***	
G9	W03	1·5	0·82	1·4				*	
G10	E08	3·1	1·09	1·6				*	
	250 ft. level								
H1	W15	16·0	2·56	2·1		***	***	**	
H2	W12	20·5	7·64	5·8			*	****	
H3	W10	6·7	2·63	2·5				***	*
H4	W49	3·5	1·50	1·9			**	***	
H5	W51	6·6	2·85	1·5			**	***	
H6	W54	3·2	1·60	1·3			*	**	
H8	W13-W24	31·1	2·94	1·8		***	**	**	*
H9	W58	2·3	0·96	1·0				***	

mineralogical investigation, therefore, the following zones may be distinguished in the mineralised clay. (Fig. 7, 8.)

(1) An upper central portion, extending to the surface, with vanadinite dominating over descloizite, the two vanadium minerals occurring almost to the exclusion of cerussite and galena. Along its western margin this zone overlaps on to a cerussite-rich zone.

(2) An intermediate zone surrounding the first on three sides and containing descloizite in excess of vanadinite.

(3) Two isolated zones flanking the first and superimposed on the second: they are rich in cerussite and are characterised by cores containing remnants of galena.

A subordinate galena-cerussite area occurs further eastward in association with willemite.

(4) An outer peripheral zone with descloizite as the only ore mineral, interrupted near the centre of the 340 ft. level by zone No. 2.

(5) A marginal willemite zone, including the near-surface Zinc Reef and extending down to the 340 ft. level along the eastern side of the ore body, where it overlaps on to zones 2, 3 and 4.

In Figs. 7 and 8 the distribution of lead and vanadium ores are, for the sake of convenience, shown separately. Minor deviations from the zonal pattern occur. A certain amount of speculation necessarily enters into the drawing of zone boundaries between levels. During the construction of the diagrams, mineral specimens collected at localities other than those of the representative samples were also taken into account. In parts of the mine where no mineralogical information could be obtained, the available assay values were used to determine the probable mineralogical composition of the ore.

The zoning of the ore body is also reflected by assay plans of the mine. Contour diagrams (Figs. 10, 11) of average lead and vanadium values were prepared on longitudinal projections of the ore body for purposes of comparison with the mineralogical diagrams described above. Unfortunately, zinc assays were only available for the Zinc Reef (Fig. 12). Values obtained by assaying the representative samples, however, (see Table II) indicate that the zinc content of the ore body itself varies between 1% and 7%, which is low compared with lead. The contour diagram for lead shows that lead is concentrated in the central part of the ore body with two peak areas coinciding with the cerussite-galena zones. They are separated by a strip with lower lead content. The contour diagram for vanadium is less regular in form but shows a well-defined peak area where the upper central vanadinite-descloizite zone is situated. A relatively rich vanadium area is indicated in the western part of the 250 ft. level, falling within the intermediate descloizite-vanadinite zone, while several isolated pockets of rich vanadium ore (descloizite) are shown on the 340 ft. level. A hypothetical area with moderate vanadium content is situated in the east between the 200 ft. and 340 ft. levels.

Used in conjunction with the contoured assay values, the mineralogical diagrams should afford a good indication of both the grade and coposition of them ore to be encountered in any part of the mine down to the 340 ft. level.

In the cerussite zones the ratio of lead to vanadium is much higher, of course, than elsewhere. A study of the assay plans reveals that this is not solely due to an increase of lead, but also to a concomitant decrease of vanadium. If the assay values on each level are plotted against the horizontal distance, lead and vanadium are found to proceed sympathetically until a cerussite-rich area is encountered. (See. Fig. 13.) Antipathetic areas coinciding with the cerussite zones are found on all levels except the 340 ft. level where the relationship is sympathetic throughout. The antipathetic rule seems to apply less well to the upper part of the western cerussite zone, the boundaries of which are somewhat uncertain. Nevertheless, it seems to indicate that cerussite was deposited in preference to vanadinite in these areas.

Calculations from Assay Values.

In the case of ore as complex and variable as that of Abenab West it is seldom possible to calculate the mineralogical composition of mine samples by means of a simple formula. If a sample consists, for instance, of galena, cerussite, vanadinite, descloizite and willemite, the mineral percentages cannot be obtained from the lead,

FIG. 7

GALENA & CERUSSITE — CERUSSITE ONLY — WILLEMITE

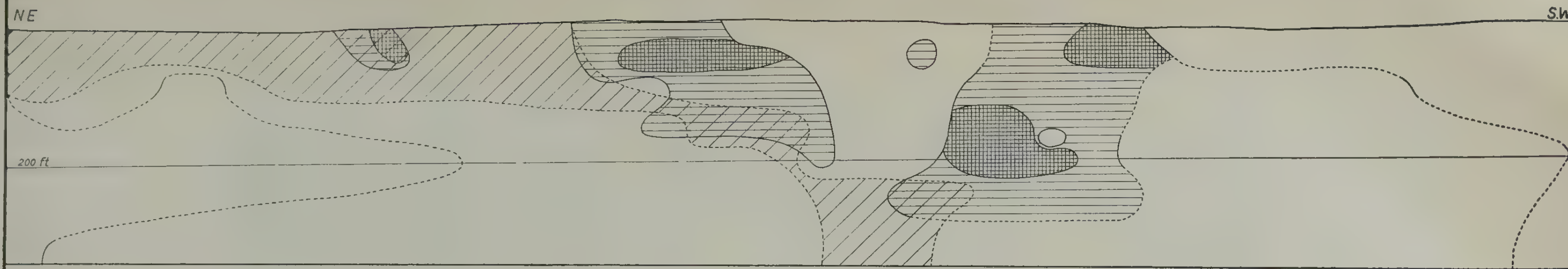
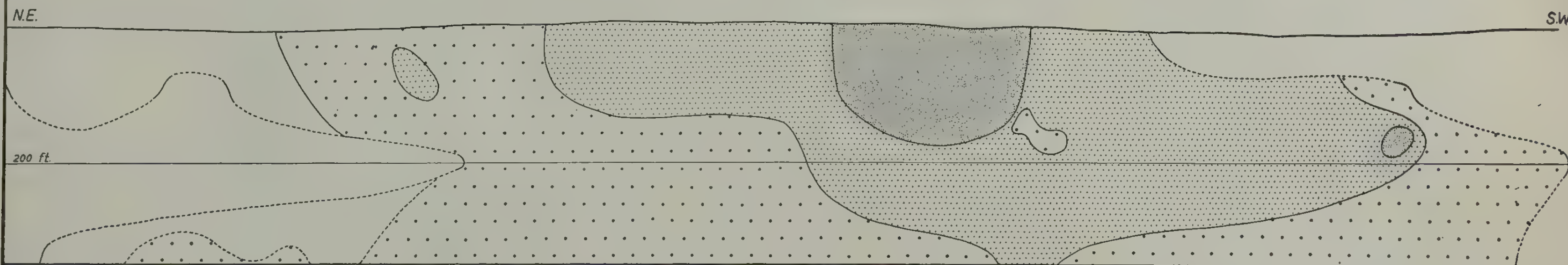


FIG. 8:

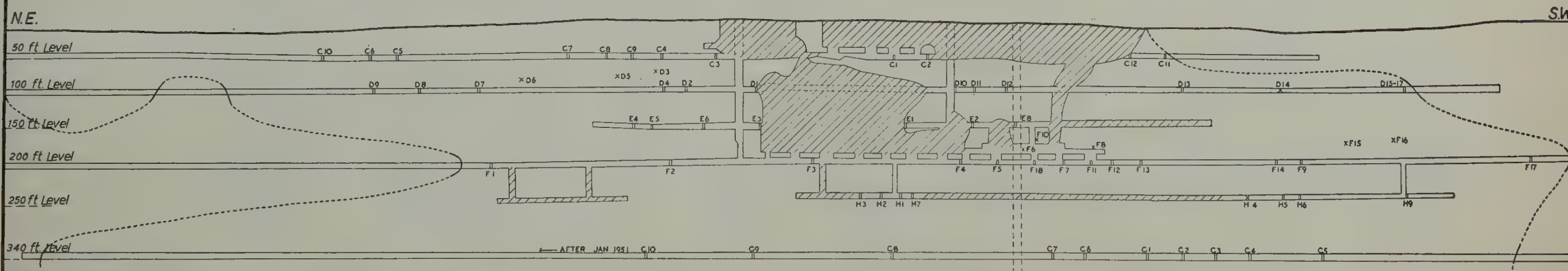
VANADINITE > DESCLOIZITE — DESCLOIZITE > & = VANADINITE — DESCLOIZITE ONLY



FIGS. 7 & 8: LONGITUDINAL PROJECTIONS OF A.W. ORE BODY SHOWING DISTRIBUTION OF ORE MINERALS

FIG. 9

LIMITS OF ORE BODY — STOPED OR INACCESSIBLE AREAS AS IN JANUARY, 1951.



SCALE 1:3000 LONGITUDINAL PROJECTION OF A.W. MINE SHOWING LOCATION OF SAMPLES

FIG. 10: LEAD

0-10% Pb · 10-20% Pb · 20-30% Pb · 30-40% Pb · >40% Pb

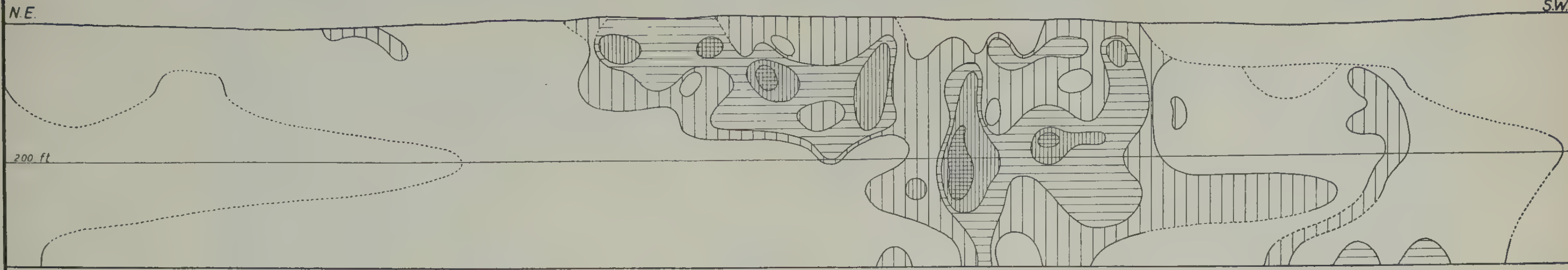


FIG. 11 VANADIUM PENTOXIDE

0-1% V_2O_5 · 1-2% V_2O_5 · 2-4% V_2O_5 · 4-6% V_2O_5 · >6% V_2O_5

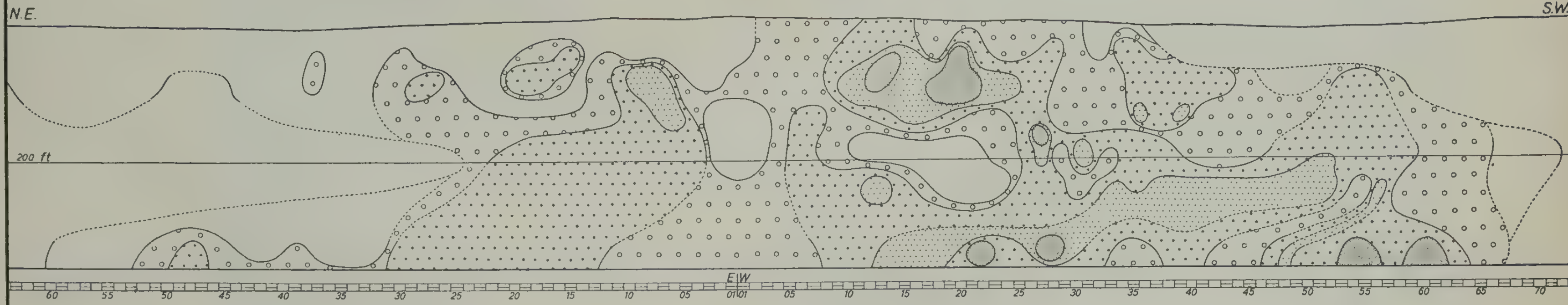
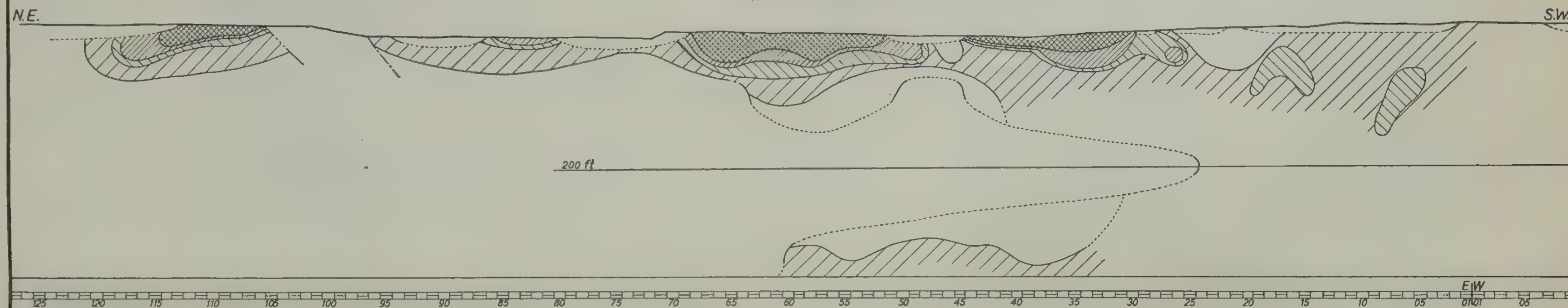


FIG. 12. ZINC

0-1% Zn · 1-5% Zn · 5-10% Zn · 10-30% Zn · >30% Zn



FIGS. 10-12: LONGITUDINAL PROJECTIONS OF A.W. ORE BODY SHOWING AVERAGE ASSAY CONTOURS SCALE 1:3000

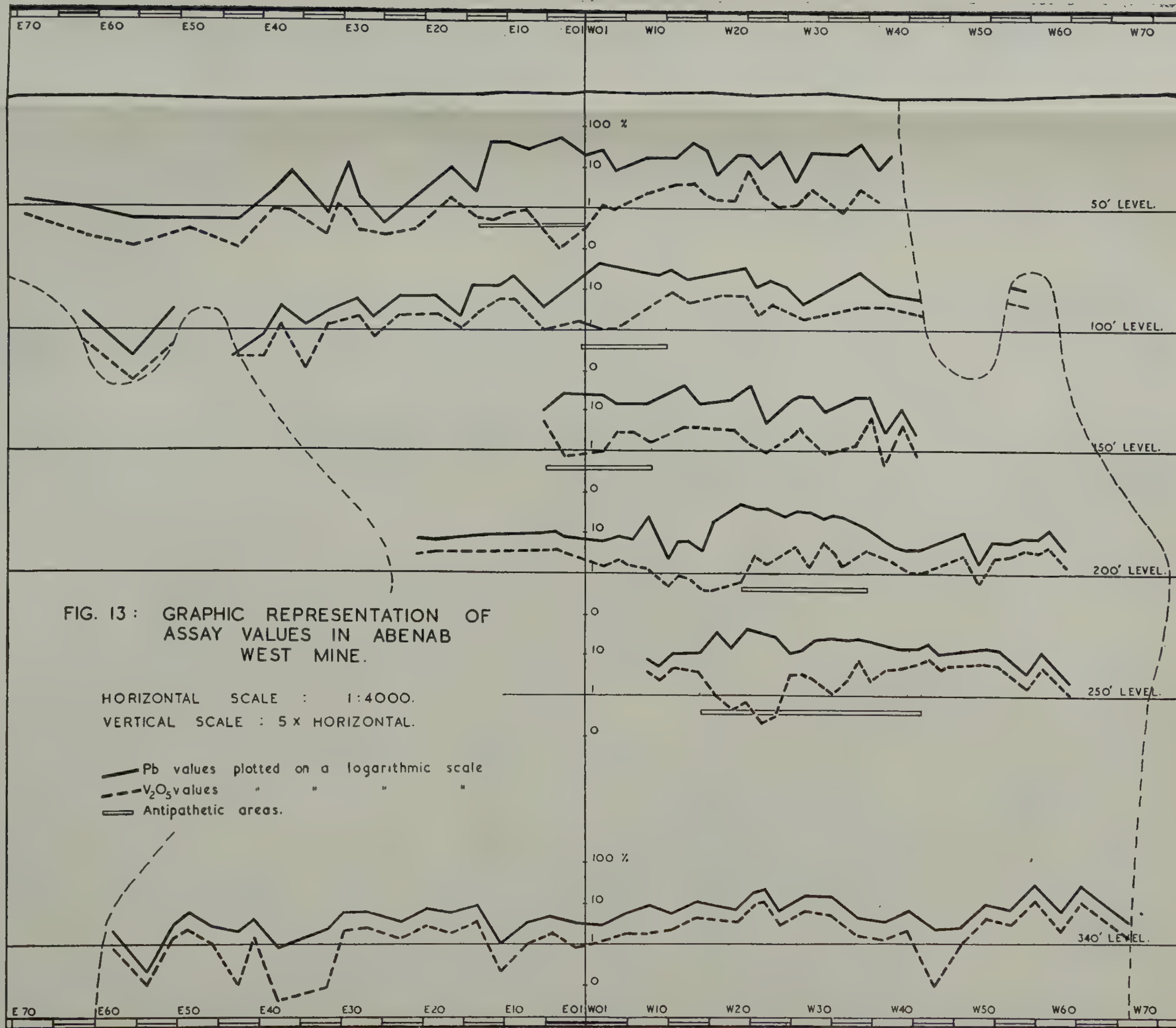


FIG. 13: GRAPHIC REPRESENTATION OF ASSAY VALUES IN ABENAB WEST MINE.

zinc and vanadium assay values only. If one of the chemical constituents is present in only one mineral, the calcuated percentage of the mineral concerned may be based on that constituent, making, perhaps, the calculation of the other mineral percentages possible. Thus, if a sample is known (or assumed) to contain only cerussite, vanadinite and descloizite, the mineralogical composition may be calculated as follows:

(1) Calculate the percentage of descloizite from the assay value for zinc.
(2) Calculate the percentage of vanadinite from the percentage of vanadium left after the allocation to descloizite.
(3) Calculate the percentage of cerussite from the percentage of lead left after allocation to both descloizite and vanadinite.

According to the microscopic investigation of the representative samples (Table II), a large number of them contain only the three ore minerals enumerated. Attempts were therefore made to calculate the mineralogical composition of these samples as outlined above. The calculation of descloizite was based on Strickland's analysis of the Abenab West mineral, recalculated to 100 (Table 3, No. 2), while for cerussite and vanadinite the theoretical compositions were used. The attempts showed, however, that in the majority of samples the zinc value yields a percentage of descloizite too high for the available V_2O_5. There appears to be an unknown proportion of zinc not accounted for by descloizite or willemite. In samples where descloizite may be assumed to be the only vanadium mineral present, this excess of zinc can be calculated by basing the descloizite percentage on the vanadium assay (Table 9).

TABLE 9: CALCULATIONS OF EXCESS ZINC IN ABENAB WEST ORE

Representative Sample	Percentage Descloizite based on V_2O_5 Assay.	Percentage Zn in Descloizite.	Zn Assay Percentage.	Excess Zn + Deficiency.
C2 (*a*)	0	0	1·1	+ 1·1
C4 (*a*)	1·4	0·2	1·0	+ 0·8
C9 (*a*)	2·8	0·4	1·8	+ 1·4
D6 (*a*)	13·9	2·1	3·4	+ 1·3
D7 (*a*)	3·6	0·6	1·5	+ 0·9
D9 (*a*)	7·5	1·2	2·8	+ 1·6
D15 (*a*)	17·2	2·6	3·5	+ 0·9
E3 (*a*)	3·4	0·5	1·3	+ 0·8
E4 (*a*)	15·2	2·3	6·9	+ 4·6
E5 (*a*)	5·2	0·8	3·0	+ 2·2
E8 (*a*)	14·4	2·2	1·1	— 1·1
F1 (*a*)	23·2	3·6	3·7	+ 0·1
F2 (*a*)	11·1	1·7	2·4	+ 0·7
F10 (*a*)	74·4	11·4	10·6	— 0·8
F17 (*a*)	2·2	0·3	1·6	+ 1·3

These figures are sufficient to show that the zinc assays are useless as a basis of calculation because a large proportion of the zinc appears to be present in an unidentified form. The validity of the assumption that V_2O_5 is only present in descloizite in the samples under consideration may be questioned; but the indications are that the vanadium is, at least in the great majority of samples, not in excess of the amount required by the other constituents to form descloizite. (For the 340 ft. level

see below.) The vanadium assays have, at any rate, limited application as a basis of calculation because most of the ore contains both descloizite and vanadinite. The lead value is likewise useless because lead is present in all the ore minerals except willemite.

By extending the number of constituents determined by chemical analysis, other ways of calculating the mineral composition become possible. The following possibilities suggest themselves:

(1) Vanadinite may be calculated from the Cl value. Because Cl is such a minor constituent it is not a very safe basis of calculation. The possible presence of mimetite renders it of doubtful value, but the amount of the latter may be judged or calculated from the As_2O_5 value and a correction applied if necessary.

(2) Descloizite may be calculated from the V_2O_5 left after allocation to vanadinite.

TABLE 10: ANALYSES OF REPRESENTATIVE SAMPLES OF ABENAB WEST ORE.

	C5 (*c*)	D10 (*c*)	F7 (*c*)	F16 (*c*)	G5 (*c*)	G10 (*c*)	Clay Fraction of D2 (*c*)
PbO	5·81	29·07	30·29	19·22	4·17	3·00	2·40
ZnO	4·35	0·09	1·88	3·24	2·09	2·74	10·21
CuO	0·06	0·10	0·08	0·15	0·06	0·09	0·11
MnO ...	0·03	0·04	0·13	0·09	0·08	0·08	0·10
Fe_2O_3... ...	28·39	30·73	13·51	51·59	12·53	40·65	24·17
Al_2O_3... ...	16·00	13·57	15·40	14·52	29·94	17·92	25·24
CaO	0·17	2·36	4·23	0·17	2·93	1·14	8·76
MgO	1·16	0·68	1·77	0·60	2·65	1·43	1·28
V_2O_5	0·70	7·82	0·98	5·72	1·79	0·91	0·83
As_2O_5... ...	0·05	0·12	0·09	0·11	0·02	0·04	0·10
CO_2	8·77	0·82	8·53	0·18	3·55	1·04	6·34
SiO_2	32·61	15·86	21·20	6·47	38·82	31·54	18·77
S	0·014	0·03	0·06	0·07	0·015	0·03	0·02
Cl	0·044	0·77	0·05	0·40	0·04	0·02	0·03
H_2O (110°C.)	1·98	0·84	1·61	0·68	2·12	1·51	2·63
Sum	100·14	102·90	99·81	103·21	100·80	102·14	100·99
Cl, S-O ...	0·02	0·19	0·04	0·13	0·02	0·02	0·02
Total ...	100·12	102·71	99·77	103·08	100·78	102·12	100·97

Note: Moisture was determined at 110°C. No information is available as to what temperature is satisfactory for the dehydration of limonite. Too high a temperature would result in CO_2 loss and destruction of organic matter. Facilities are not available for the determination of ferrous iron in the presence of ferric iron. All Fe has been calculated to Fe_2O_3.

E. H. STRICKLAND (*Analyst*),
Abenab Assay Office.

(3) Willemite may be calculated from the Zn left after the calculation of descloizite. The result would be questionable due to the presence of unaccounted Zn. Willemite cannot be calculated from SiO_2 because unknown amounts of quartz and clay silicates are present.

(4) Galena may be calculated from S. Galena seems to be the only sulphide of importance, but the determination of very small amounts of S to the required degree of accuracy would be difficult.
(5) Cerussite may be calculated from the Pb left after the calculation of vanadinite, descloizite and galena.
(6) Cerussite may also be calculated from the CO_2 left after satisfaction of Ca and Mg to form calcite and dolomite. This calculation is based on the assumption, among others, that all Mg is present in dolomite, which is probably not true in view of the presence of montmorillonite, which has a highly variable composition.

Attempts to calculate the percentages of the ore minerals in a few samples from their complete analyses (Table 10) along these lines were not very successful owing to the difficulties noted above. The ore seems to be of too complex a nature for mineralogical calculations of this kind.

The agreement between assay values and mineralogical observations is not perfect even in the case of ore with an apparently simple composition. For example, while calculating the percentages of descloizite in samples from the 340 ft. level containing descloizite as the only ore mineral, zinc was found to be in slight excess in some cases and deficient in others. If the relative proportions of the three constituents Pb, Zn and V_2O_5 in these samples are compared with those of pure descloizite from the same environment, all three constituents are seen to vary above and below the ideal values (Table 11).

TABLE 11: RATIO OF CONSTITUENTS OF ORE FROM THE 340-FT. LEVEL.

Sample	$Pb + Zn + V_2O_5$	$\frac{Pb \times 100}{Total}$	$\frac{Zn \times 100}{Total}$	$\frac{V_2O_5 \times 100}{Total}$
A. W. Descloizite (Strickland)	87·8	58·5	17·3	24·2
G1 (*a*)	13·95	59·5	14·3	26·2
G2 (*a*)	19·29	57·5	17·7	24·8
G3 (*a*)	12·07	57·2	18·2	24·6
G4 (*a*)	5·10	60·8	11·8	27·4
G5 (*a*)	7·00	54·3	18·6	27·1
G6 (*a*)	10·31	53·3	20·4	26·3
G8 (*a*)	28·32	57·6	17·3	25·1
G9 (*a*)	3·72	40·3	37·6	22·1
G10 (*a*)	5·79	53·5	27·7	18·8

This state of affairs may be due to (*a*) marked variations in the composition of descloizite from the same level of the mine, or (*b*) undue reliance on the assay values, or (*c*) the occurrence of two or all three chemical constituents in other mineralogical forms in the ore. The third possibility seems to be the most likely cause: the occurrence of unidentified zinc in other parts of the mine has already been demonstrated, while lead and zinc-bearing calcite is known to occur on the 340 ft. level. The amounts of lead and zinc in the calcite (See Table 4) seem to be insufficient to be solely responsible for the discrepancies. The possibility that a small percentage of vanadium occurs adsorbed in the clay merits further investigation.

The failure of the calculation methods induced the author to use the lead/zinc/vanadium ratios of assayed samples as a basis for distinguishing between different types of ore without previous mineralogic study. The relative proportions of the three chemical constituents were plotted on a triangular diagram, but it was found that the vanadinite, descloizite and cerussite-galena fields overlapped to such an extent that distinction was impossible. The lead/vanadium ratios of descloizite and vanadinite are comparatively close together while the possible presence of willemite and unidentified zinc makes the determination of mineralogical composition without a qualitative knowledge of the minerals present an impossible task.

The Excess Zinc.

So far zinc is the only constituent that could be definitely shown to occur consistently in a form unaccounted for by the minerals identified. There is no reason to believe that the assay values for zinc can be several per cent too high. The following possible sources for the excess of zinc may be suggested:

(1) Zincian calcite. As noted above, a content of 0·03 to 0·4 Zn in the pure mineral is not sufficient to explain the discrepancies of 1-4% in the composite ore.

(2) An earthy zinc oxide or silicate.

(3) A zinc-bearing clay mineral.

In this connection it is of interest that zinciferous clay is an important constituent of the oxidized zinc ores of Leadville, Colorado (Loughlin, 1918). This clay occurs as relatively pure layers large enough to be called small ore bodies. There is an inverse relationship between the zinc oxide and alumina content of the clay, suggesting the isomorphous replacement of alumina by zinc. On the basis of microscopic evidence, Loughlin concluded that the clay consists of "kaolin containing a small percentage of zinc, mostly in the form of a sericite-like mineral."

In an attempt to determine the chemical composition of the unknown zinc mineral at Abenab West, the clay fraction of a sample containing a large proportion of excess zinc (D2 (*c*)) was separated by decantation and submitted for complete analysis. The result (see Table 13) showed that the excess zinc is definitely concentrated within the clay. If the theoretical mineralogical composition of the sample is calculated as explained in previous pages, it is found to contain 9·7% ZnO in excess of descloizite together with 1·2% vanadinite, 2·8% descloizite and an unknown percentage of MgO in excess of the requirements for dolomite. The chemical combination of the silica, alumina and ferric oxide is too uncertain to warrant further conclusions, though it seems quite possible that the zinc may be present in an aluminosilicate. No direct relationship between zinc oxide and alumina or any other constituent is revealed by arranging the available analyses of composite Abenab West ore in the order of increasing alumina. The Abenab West clay is too much contaminated with ferric oxide to make microscopic observations on the clay minerals possible. Even the X-ray examination of the clay samples provided no clue to the identity of the unknown zinc mineral.

Until relatively pure samples of zinciferous clay are found in the Abenab West mine, the excess zinc in the ore will probably remain unexplained.

Conclusions.

(1) The Abenab West ore body consists of three concentric vanadinite-descloizite zones with vanadinite predominating in the central one and descloizite increasing towards the periphery. These are superimposed on a twofold galena-

cerussite core and the whole is flanked on the east by willemite rock concentrated especially at the outcrop.

(2) The approximate mineralogical composition and grade of the ore in any part of the mine may be determined by using the mineralogical diagrams in conjunction with assay plans.

(3) The approximate mineralogical composition of any ore sample can be determined microscopically by estimation of the relative amounts of the ore minerals in a panning concentrate.

(4) If a suitable method of concentration can be developed, the quantitative mineralogical composition of ore samples can be determined microscopically by a trained mineralogist with the aid of grain counting, provided the mineral grains are correctly identified.

(5) In ore as complex and variable as that of Abenab West it is unlikely that the mineralogical composition of samples can be determined simply by calculation from assay values or analyses.

(6) Excess of zinc unaccounted for by the known zinc minerals is present above the 340 ft. level, probably in the form of zinciferous clay.

No confirmation could be found for the belief, current at the mine, that as much as 10% lead in excess of the composition of the known lead minerals is present in the ore. This assumption is apparently based on an erroneous calculation in which the composition of vanadinite was incorrectly applied.

(7) As a result of uncertainties that still exist concerning the mineralogical distribution of some chemical constituents in the ore, the correspondence between the assay values and the results of mineral determinations is not satisfactory.

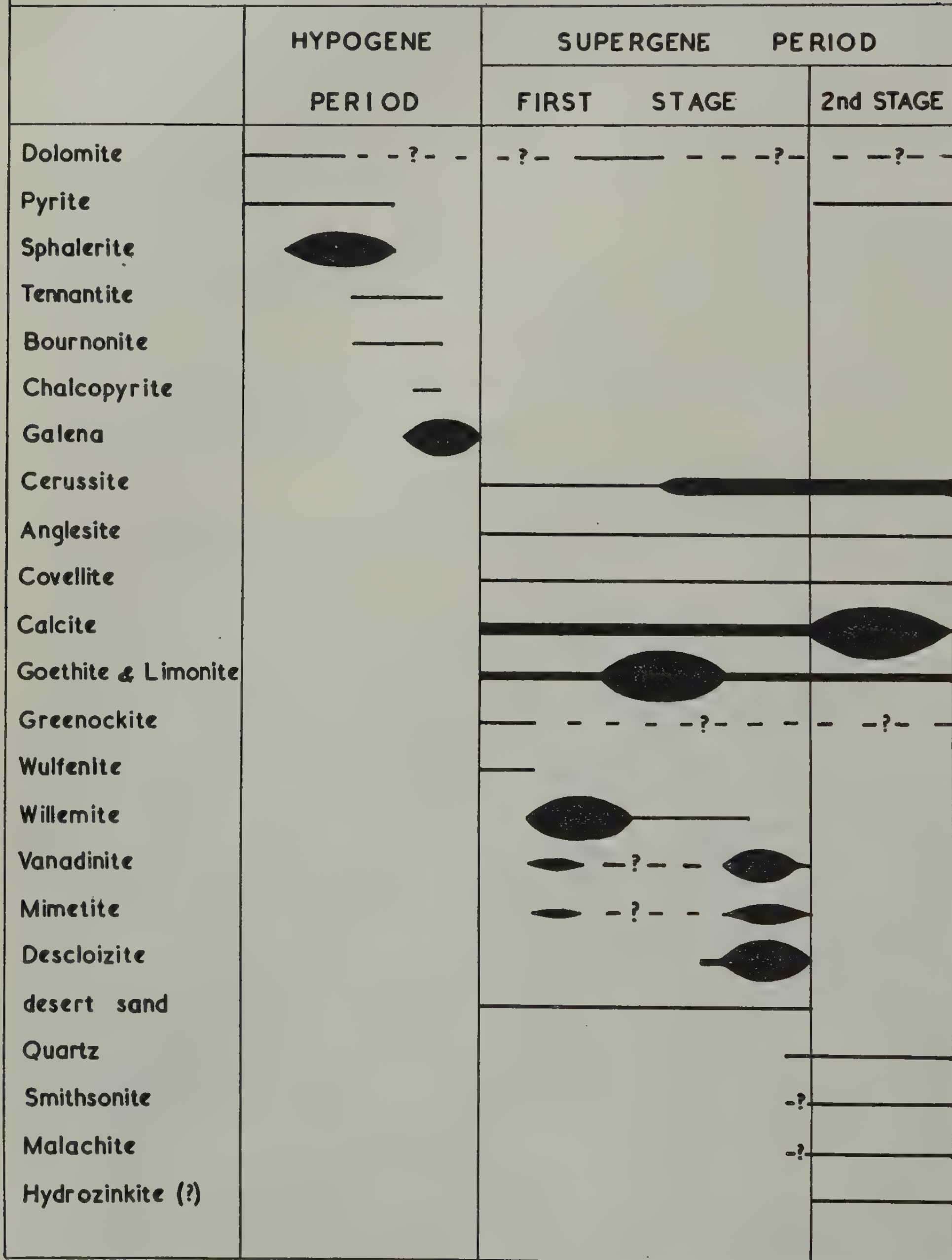
IX PARAGENESIS AND GEOLOGIC HISTORY OF THE ABENAB WEST DEPOSIT

From the foregoing pages it should be clear that the Abenab West ore body is considered to be the oxidized and contaminated hood zone of a sulphide vein. The principal evidence for this view is: (1) the presence of massive sulphide ore at least four and a half feet thick, along the same horizon about 300 feet below the deepest exposure of mineralised clay, (2) the zonal distribution of the secondary minerals, producing a lead-rich core and a zinc-vanadium-rich periphery, and (3) the occurrence of partly oxidized sulphide ore in the clay. The size of these inclusions militates against the view that they were merely incorporated from the disseminated sulphides in the wall rock. While the latter may be measured in centimetres at most, some of the galena blocks in the clay weigh a few hundred pounds.

Thus the paragenetic history of the ore body may be divided into (i) a hypogene and (ii) a supergene period. The supergene period was characterised by a series of transformations involving the repeated solution, deposition and replacement of secondary minerals like willemite, vanadinite, mimetite and calcite under the influence of changing acidity, concentration, composition and temperature of ground waters. Two stages of oxidation are distinguished below. It is certain that contamination from outside played a part in the formation of the ore body and it is of considerable interest to know the position of the vanadium mineralisation in the sequence of events.

In Fig. 14 the sequence of deposition of all the minerals at Abenab West is shown diagramatically. (Time divisions and vertical dimensions do not represent true proportions.)

FIG. 14 : PARAGENETIC DIAGRAM



Hypogene Period

The primary ore minerals in the massive vein of drill core no. 23 and in the oxidized clay are sphalerite, galena, pyrite, tennantite, bournonite and chalcopyrite, more or less in the order of abundance. From a study of mineral contacts it seems as if pyrite and sphalerite were the first to crystallise; some galena and most tennantite were caught in the sphalerite mesh, but some tennantite was squeezed out to crystallise on the outer margins, followed by the consolidation of galena with its bournonite and chalcopyrite inclusions.

The bulk composition of the original sulphide ore may be deduced to some extent form the composition of the oxidized zone. The scarcity of cupriferous secondary minerals, for instance, is in full agreement with the low copper content of the sulphide remnants. Lack of copper and abundant pyrite in the former sulphides are also indicated by the occurrence of transported instead of indigenous limonite and by the brilliant red and yellow colours of the ferric hydrates. (Bateman, 1942.) This is in marked contrast with the Tsumeb ore body, which is so poor in iron that pyrite must be added during smelting operations (Schneiderhühn, 1929). Nevertheless, the primary mineral assemblage at Abenab West is not dissimilar to that of the lower levels of the Tsumeb mine, where copper minerals are rather subordinate to galena and sphalerite. It is possible that the Abenab West ore body represents only the root portion of an originally much larger deposit.

There is undoubtedly a genetic relationship between the former sulphide mass at Abenab West and the disseminated sulphide blebs in the country rock. They are restricted to the same stratigraphic horizon and have the same mineral constituents. The close correspondence in the distribution of lead and zinc and in the positions of minimum and maximum mineralisation clearly indicates that the massive and disseminated sulphides are contemporaneous. The disseminated sulphides crystallised mainly in small cavities, probably opened up in advance by mineralising solutions which dissolved the dolomitic country rock. Part of the dissolved material was ultimately also precipitated as clusters of white dolomite crystals. The texture of the sulphide mass in drill core no. 23 also seems to indicate crystallisation in an open space rather than replacement of country rock.

The sulphide ores of Abenab West and Tsumeb belong to the same epoch of mineralisation and similar conditions of deposition may be assumed. The presence of sulpharsenides and sulphantimonides (eg. tennantite and bournonite at the former and tennantite, enargite, luzonite, germanite at the latter locality) is indicative of a mesothermal origin. The temperature of formation is also reflected by some of the trace elements in sphalerite, but lack of similarity in sphalerite from different metallogenetic regions must be taken into account. (Stoiber, 1940; Oftedal, 1940; Rankama and Sahama, 1950.) In general, In favours sphalerites formed at intermediate temperatures and Ga and Ge decrease in successively higher temperature deposits whereas Mn increases. At Tsumeb the Ga and Ge percentages are in agreement with an intermediate temperature of formation but In is a bit low. (Moritz, 1933; Oftedal, 1940.) At Abenab West the content of trace elements is only known qualitatively and does not allow definite conclusions. Fe predominates over Mn and Ge in sphalerite, In is absent and Ga is apparently found in galena instead of the more usual sphalerite.

The emplacement of the sulphides at Abenab West was followed by tectonic movements that are largely responsible for the present shape of the ore body. The evidence that at least the minor faulting is later than the sulphides is as follows:

(i) The arcuate faults truncate the disseminated sulphide zone which is considered to be contemporaneous with the main sulphide vein.

(ii) Remnants of galena show a foliation structure and variable grain size due to recrystallisation which is most probably related to localised deformation.

(iii) The disturbances reach their maximum at the surface and may be interpreted as relatively recent phenomena whereas the sulphide body is thought to have experienced considerable denudation.

The faults are closely related to the small folds or "rolls" affecting the ore body. It may be suggested that these, too, are subsequent and did not localise the emplacement of the sulphides. The incompetency of the sulphide vein during deformation resulted in a thickening along the crest and a thinning along the limbs of a fold. Such a highly disturbed sulphide vein may be expected to give easy access to oxidizing ground waters. The movements therefore provide a partial explanation for the remarkably complete oxidation that followed.

Supergene Period: First Stage.

The general rule that sphalerite is the first of the sulphides to be attacked in the zone of oxidation is well demonstrated at Abenab West. In the drill core and in sulphide blebs partly altered sphalerite is seen in contact with perfectly fresh galena and pyrite. In the oxidized ore body sphalerite has been totally removed except for small inclusions in galena. Whereas galena is surrounded by a protective covering of sparingly soluble lead carbonate and sulphate, the soluble zinc salts have migrated considerable distances before precipitation as shown by the peripheral relationship of the secondary zinc minerals at Abenab West.

The chemical changes involved in the decomposition of the sulphide ore were probably based on the following reactions. Initially, carbonic acid and oxygen were the only active reagents and the following equilibria were established:

$$H_2CO_3 \rightleftharpoons H^+ + HCO_3^- \qquad HCO_3^- \rightleftharpoons H^+ + CO_3^{=}$$

$$ZnS + 2H^+ \rightleftharpoons Zn^{++} + H_2S$$

$$H_2S + O_2 \rightleftharpoons 2H^+ + SO_4^{=}$$

Owing to the weak acidity of the waters, oxidation proceeded slowly in the early stages. Experiments by Gottschalk and Bühler (1910, 1912) have shown that in mixed sulphides the oxidation and solution of sphalerite (with a low potential) is strongly accelerated by electrolytic action whilst retarding the oxidation of pyrite (with a high potential). Though pyrite was thus attacked much more slowly than sphalerite, the oxidation of even small amounts of pyrite enormously enhanced the oxidation of all the sulphides by the liberation of sulphuric acid and the powerful solvent ferric sulphate. (See Bateman, 1942.) For example, with sphalerite the reaction may be expressed as:

$$ZnS + 4Fe_2(SO_4)_3 = ZnSO_4 + 8FeSO_4 + 4H_2SO_4$$

The oxidation of sphalerite was therefore accompanied by the simultaneous oxidation of some pyrite and a little galena, but large amounts of the other two sulphides remained after the practically complete removal of sphalerite. Zinc and iron were kept in solution by $SO_4^{=}$ and to some extent by HCO_3^- and Cl^-. Lead, however, was quickly precipitated as anglesite and cerussite, or managed to remain in solution and was transported a short distance from its source before deposition. Traces of copper in the galena and sphalerite were liberated during oxidation, augmented by the decomposition of bournonite and tennantite and were deposited as covellite. Similarly, the small amounts of cadmium in the crystal lattice of sphalerite were separated from the zinc and remained behind as greenockite while the zinc was removed in solution.

Deposition of the Zinc Reef: There are three possible sources of the zinc in the Zinc Reef: (*a*) the original sulphide vein to the west of the Zinc Reef, where the present ore body is situated, (*b*) a since eroded portion of the sulphide vein, formerly situated above the Zinc Reef, and (*c*) the sphalerite blebs in the Zinc Reef itself. Upward and lateral migration of solutions from the first, and possibly downward percolation from the second source are considered to have been the principal modes of derivation, while the sphalerite blebs must have yield an insignificant contribution.

The solutions circulating from the sulphides into the country rock were definitely acid and contained at least the following ions: Zn^{2+}, Fe^{3+}, Ca^{2+}, H^+, $SO_4^=$, $CO_3^=$, HCO_3^-. The fact that willemite was deposited shows that silica was an important constituent too, probably in the form of a hydrated cation. (The origin of the silica is not obvious, but it may have been derived from the weathering of shale in the vicinity of the ore body.) The conditions of chemical equilibrium that exist when so complex a solution attacks limestone and dolomite can only be roughly inferred from experimental data and mineralogical observations. Lack of solubility is the predominating factor in determining the order of separation of the products of such a reaction. (Loughlin, 1918.)

Where the solution came into contact with the carbonate rock, the excess of sulphuric acid would at once be neutralised by reaction with the limestone and dolomite,—Ca^{2+}, Mg^{2+} and $SO_4^=$ passing off in solution. Ferric sulphate would react further with the carbonate rock and be precipitated as ferric hydroxide according to the following equation:

$$3CaCO_3 + 2Fe^{3+} + 3H_2O = 2Fe(OH)_3 + 3Ca^{2+} + 3CO_2.$$

The ferric hydroxide formed would, by gradual loss of part of its water, be gradually transformed into one of the hydrated oxides and thus account for the red colouration in the Zinc Reef and vicinity.

At this stage, silica must have been the dominant acid radicle because the indications are that zinc sulphate and bicarbonate could no longer remain in solution in the presence of excess of silica. Willemite started to crystallise by metasomatic interchange with the country rock, or by deposition in spaces provided by the acid attack. The siliceous waters penetrated sphalerite blebs along cleavage directions and reacted with them to form willemite veinlets in place. Greenockite, the secondary cadmium sulphide, was deposited in direct contact with this sphalerite, which probably acted as reducing agent. Apparently, secondary galena(?) was deposited by a similar cementation process. Small quantities of wulfenite crystallised simultaneously with willemite. Molybdenum has not been detected in any other mineral at Abenab West, but the amount involved is so small that there is not much difficulty in ascribing it to its traditional source—the primary sulphides from which the lead and zinc were derived. Any excess of zinc or silica left after the deposition of willemite was removed in solution and it is possible that this zinc was utilised somewhat later in the formation of descloizite.

It is a remarkable fact that no smithsonite was deposited at Abenab West before or after the crystallisation of willemite. Though the oxidizing solutions circulated in a highly calcareous environment, CO_2 appears to have been a very subordinate constituent to SiO_2. This state of affairs might have obtained if the solutions had been concentrated. In contrast with the situation at Abenab West, smithsonite is the most important secondary zinc mineral at the analogous lead-zinc-vanadium deposit at Berg Aukas and in the oxidized zone at Tsumeb. Nevertheless, willemite is known to occur at several other localities in the mineral district, though it may differ somewhat from the Abenab West type. L. J. Spencer (1927) described a

specimen from Guchab, near Otavi, as showing stout hexagonal prisms, 2·5 to 5 mm. across, sulphur-yellow on the exterior but white and grey inside, occurring with botryoidal malachite on massive chalcocite. In mineral specimens from Berg Aukas, spheroidal aggregates and drusy crystals of willemite were found to be quite common and earlier than smithsonite. According to a verbal communication by Dr. Brandt a "Zinc Reef" similar to the one at Abenab West is situated near the old Baltica vanadium mine. Many willemite localities have probably been hitherto overlooked because it is not a very conspicuous or valuable mineral. Willemite is definitely of widespread occurrence in the Otavi Mountain Land and appears to be an associate of the vanadium ores. It is clear that the conditions necessary for the formation of willemite must have prevailed throughout the district.

Where silica is present in weathering solutions, the normal oxidation product of sphalerite is the hydrous zinc silicate hemimorphite, and not willemite. To the best of the author's knowledge, hemimorphite has not been found in South West Africa. The fact that willemite was deposited instead of hemimorphite thus requires an explanation. It must be emphasised that willemite, though uncommon, is typically found in the oxidized zone of ore deposits and that its occurrence in hypogene form, as in the unique deposit at Franklin, New Jersey, is mineralogically distinct and exceptional. To hypogene willemite the Abenab West mineral bears no resemblance, while in mode of occurrence, crystal form, chemical composition and lack of fluorescence it is closely analogous to supergene willemite from such localities as Chihuahua, Mexico; Hillsboro, Arizona; and Sobral da Adiça, Portugal. (Pough, 1941.)

The problem of the conditions necessary for the formation of supergene willemite has received the attention of F. H. Pough (1941), who noted that with two exceptions all known occurrences are located in arid regions. He concludes that the most important factor in favour of the deposition of willemite is high concentration—a condition that is true in weathering solutions of arid climates, but which may occur locally elsewhere. Other possible factors are relative acidity and high pressure and temperature, which may be brought about by oxidation to considerable depths if the ground water table is low.

The Climate

Though bounded by desert regions in the west and in a south-easterly direction, the Otavi Mountain Land has by no means an arid climate. According to the rainfall map in the annual Weather Bureau reports for South West Africa (based on Zelle's map of normal isohytes for the 35-year period 1901-1936) the area enclosing Otavi, Tsumeb, Tsintsabis and Grootfontein enjoys a normal annual precipitation of 20-25 inches (50·8-63·5 cm.). The mean climatological data for the metereological station at Tsumeb (Fig. 15) may be taken as representative for the area in question. These have been compiled from all available data over the period 1912-1950 by the staff of the Weather Office, Windhoek, to whom the author wishes to express his gratitude. The mean annual precipitation at this station is 52·5 cm., the mean annual temperature is 22·1°C. and the figures for relative humidity indicate moderate evaporation, in the summer at least, and a correspondingly high ground water content.

Though Köppen includes the Otavi Mountain Land in his warm temperate rainfall region with the climatic formula Cwa, the figures quoted above show that it should be classified as a hot steppe climate with summer rainfall, and should receive the formula BShw. (Köppen, 1931). B. R. Schulze (S.A. Geogr. J. XXIX, 1947) is in agreement with this and has furthermore applied the Thornthwaite classification to

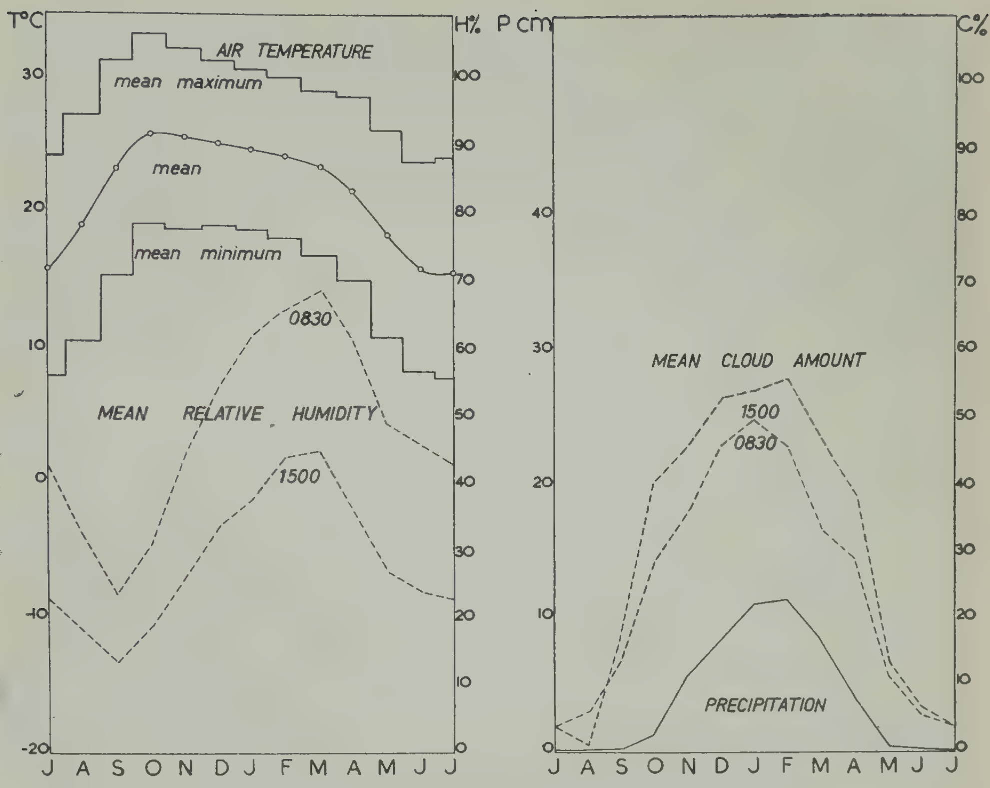


FIG. 15. Climatologic Data for Tsumeb.

the climates of South Africa, including the area under consideration in his DB¹d or semi-arid warm region. In his regional description of South African climates, S. P. Jackson (S.A. Geogr. J. XXXIII, 1951) also applies the term semi-arid to the area in question. Thus the present climate of the Otavi Mountain Land does not seem to accord very well with the postulated conditions for the deposition of secondary willemite.

There are several indications, however, of a sharp climatic change in geologically recent times. A. Penck (quoted by Rogers, 1922) argued that during the Pleistocene Period the desert belts of the earth must have been nearer the equator. A. W. Rogers (1922) finds evidence of a post-Pleistocene southward shift of the dry belt of South Africa in the distribution of pans and the fact that the sand region of the Kalahari is now covered with vegetation.

In northern South West Africa extensive deposits of red eolian sand cover the older formations and encroach in the valleys between the Otavi Mountains; but as in the Kalahari, these deposits are tied down and the prevailing winds no longer carry sand. H. Schneiderhöhn (1920) has made a detailed study of this sand and reaches the following conclusion:

"Der Klima, bei dem diese Flugsande sich ablagerten, ist ein trockenes Steppenklima anzunehmen, so wie es heute etwa im mittleren Hereroland herscht, wo die Vegetation so zurücktritt, dasz der Wind ungehindert die lockeren Sandmassen bewegen kann, wo die geringeren Niederschläge nicht hinreichen so viel Eisenhydroxyd und Kalk aufzulösen um den Sand zu verkitten." Ferric hydroxide coating around the sand grains are not present in the Namib desert and Hereroland but make their appearance where the rainfall exceeds 38 centimetres. They are therefore ascribed to a more humid period transitional to the present climate of the Otavi Mountain region.

Furthermore, the secondary mineral assemblage of the Otavi Mountains as a whole shows a remarkable correspondence with that of arid regions like the southwestern states of America. Willemite, wulfenite, descloizite, vanadinite and the rare copper silicates dioptase and plancheite occurring in the Otavi Valley (Schneiderhöhn, 1929) are all regarded as characteristic of weathering under arid conditions (Pough, 1941). The extraordinary variety and beauty of the secondary minerals at Tsumeb may also be ascribed in part to this agency.

To these arguments in favour of a former more arid period in the Otavi Mountain area may be added the evidence afforded by the Abenab West ore body itself. The oxidized ferruginous clay extends to a depth of more than 340 ft. whereas the local water table lies at approximately 250 ft. as shown by the depth of water in the adjacent Abenab open-caste and by underground streams in the Abenab West mine. The present ore body therefore has the appearance of a drowned oxidized zone, formed at a time when the ground water level stood much lower than at present. This suggestion must be regarded with caution, however, since ground water circulation may extend to great depths along a disturbed zone, especially in non-porous calcareous rocks where a regional water table does not exist.

From the above it seems to be fairly well established that the Otavi Mountain Land has experienced a period of relative aridity in the not too distant past. It is a noteworthy fact that the eolian sand grains in the Abenab West clay are not coated with ferric hydroxide as are those of the surface deposits in the vicinity. This proves that the sand was incorporated before the onset of humid conditions and hence that the oxidation of the ore body largely took place during the arid period. In the opinion of the author this fact clears up several difficulties in connection with the Abenab West ore body. It helps to explain the deep and extensive alteration the sulphides have undergone, accounts for the presence of sand grains of eolian origin in the oxidized clay and provides the reason why willemite and not hemimorphite was the zinc silicate formed. The conclusion that willemite was deposited under different climatic conditions than the present is supported by casts and replacement remnants which tend to show that willemite has become unstable since its deposition. The recognition of a semi-arid oxidation period is also thought to have an important bearing on the vanadium problem.

The Abenab West Clay

The Abenab West clay may be interpreted (*a*) as fault gouge. Since the faulting has been shown to be post-disseminated sulphides and probably post-massive sulphides in age, this source may have contributed in part. (*b*) As the detritus derived from the weathering of shale and the insoluble residue left by solution of dolomite and limestone. The presence of bedded cave deposits, large and small, in the ore body itself, proves that the deposition of transported material by ground water played an important part in the history of the ore body. The major portion of the gangue is considered to be such ferruginous "clay fill". (*c*) As the product of hydrothermal alteration. The

nature of the clay minerals and the predominance of iron oxides testify against this theory.

The introduction of foreign material must have proceeded hand in hand with the oxidation of the sulphides, and has probably continued up to the present. Alumina, ferric hydroxide and sand were the principal constituents derived in this manner. Most of the hydrated ferric oxides, however, with their brilliant red and yellow colours, colloform structure and crystallisation as goethite, may be ascribed to the oxidation of pyrite. It is evident that this "limonite" is not of the indigenous kind: no ferruginous boxworks that can be described as due to the replacement of a sulphide along cleavage directions have been found. The iron was transported in solution and perhaps partially in colloidal suspension before being deposited at some distance from its source. The bulk of the ferric hydrates were precipitated before the crystallisation of descloizite.

The sand grains in the clay were carried to the outcrop by prevailing winds during this period and washed downwards by the occasional downpours characteristic of an arid climate. There is little doubt that the golden brown mica, muscovite, microcline and the heavy mineral grains tourmaline (green, brown and blue), magnetite, zircon, rutile, epidote, kyanite, allanite and anatase are detrital constituents of the wind-blown sand in the clay. They may be compared with the following minerals listed by Schneiderhöhn (1920, p. 279) as constituents of red eolian sand from all over the Otavi Mountain Land:

Quartz (95%) Feldspars (especially microcline, 2%), sericite, tourmaline, zircon, magnetite, rutile, pistacite (epidote), clinozoisite, piedmontite, orthite (allanite), disthene (kyanite), andalusite, anatase, titanite, biotite, corundum, garnet, topaz, sillimanite, picotite, serendibite and two unidentified minerals.

The large variety of foreign minerals, the uniform grain size and the rounding of the quartz grains indicate an eolian origin. The provenance of the sand may be situated far outside the borders of South West Africa. A highly metamorphic terrain and acid igneous rocks with associated pegmatites, such as the Archean Complex of Africa, is indicated. (Schneiderhöhn, 1920.)

The genesis of the colourless tourmaline prisms in the clay is not so obvious and a wholly satisfactory explanation of its presence has not been found. According to P. D. Krynine (1946), very small colourless tourmaline crystals are found in pegmatised, injected metamorphic terrains consisting especially of phyllites. However, none of the tourmalines illustrated or described by Krynine resemble the slender prisms of Abenab West. The exceptionally low refractive indices of the Abenab West mineral ($\epsilon = 1 \cdot 61$, $\omega = 1 \cdot 63$) correspond closely to the refractive indices ($\epsilon = 1 \cdot 610$, $1 \cdot 612$; $\omega = 1 \cdot 628$, $1 \cdot 630$) of sedimentary authigenic tourmaline "formed at the bottom of the sea penecontemporaneously with the including sediment." (Krynine, 1946). This authigenic tourmaline is described and shown as colourless overgrowths on cores of various types of detrital tourmaline. The possibility that the Abenab West prisms have an authigenic origin—whether originally attached to cores of tourmaline or formed in some other fashion under cold water conditions—should not be disregarded until further evidence is forthcoming. In the meantime the tendency is strong to regard this tourmaline as primary and hypogene—perhaps a gangue mineral associated with the original sulphides.

The Vanadium Mineralisation

Observations bearing on the conditions of deposition of descloizite and vanadinite are given in the next chapter. At this stage it is merely intended to describe their positions in the paragenetic sequence.

Whatever the origin of the vanadium, it is clear that the crystallisation of the two vanadium minerals took place during the supergene stage in the history of the Abenab West ore body. The disseminated crystals were not incorporated from elsewhere, but crystallised *in situ*, and are younger than the clay. The vanadium mineralisation is therefore separated from the sulphide mineralisation by a long time interval embracing the deformation, partial oxidation and contamination of the sulphide vein.

Vanadinite was the first vanadium mineral deposited. It has been shown that there were two stages of vanadinite deposition, the earlier one antedating and the later stage post-dating most limonite. The coarsely crystalline early vanadinite has been almost completely dissolved or replaced leaving remnants, pseudomorphs and casts as indications of its former presence. These are too rare, however, to warrant the conclusion that all later vanadinite was derived from the earlier generation. The solution and interrupted deposition of vanadinite may be ascribed to normal changes in the composition of the oxidizing solutions during the secondary alteration of the ore body. This does not exclude the possibility that appreciable early vanadinite may have survived in some areas of the mine, perhaps on levels lower than 340 ft.

The depositional history of vanadinite is closely analogous to that of mimetite, with which it is isomorphous. Though rarer than vanadinite, remnants of mimetite earlier than limonite and crystals later than the same limonite have been found. Mimetite is most probably secondary after tennantite, the only primary mineral yet encountered with arsenic as a constituent. There is no doubt that the mimetite belongs to the oxidation period in the history of the ore body.

Descloizite is later than willemite, early vanadinite and limonite-goethite. In the Zinc Reef it was one of the last minerals deposited and on the 340 ft. level it crystallised simultaneously with calcite that replaces willemite. A small breccia body adjacent to the Zinc Reef contains limestone fragments, impregnated from their margins inwards with red iron oxide, and cemented by coarse calcite and descloizite, just like the old Abenab ore. This occurrence confirms the conclusion that the calcite-descloizite association post-dates the deposition of willemite, which was presumably related to the ferruginisation of the brecciated limestone. Descloizite is also found to replace coarsely crystalline early vanadinite. There is no direct evidence of the relationship between descloizite and late vanadinite, but they may be considered as contemporaneous. Small drusy crystals of both minerals encrust porous limonite and are disseminated in the clay, the vanadinite being more or less restricted to the upper central portions of the ore body while the descloizite forms a zinc-rich periphery.

It is remarkable that vanadinite and not descloizite should be the first vanadium mineral formed. If the heavy metal content of the vanadium minerals is considered to be derived from the primary sulphides, descloizite should be earlier than vanadinite because zinc is more easily mobilised than lead. That vanadium was present in solution during the deposition of willemite is proved by the small percentage of vanadium in wulfenite. The explanation here suggested is that the conditions (pH?) were not suitable for the deposition of descloizite until willemite had crystallised.

The geological relationship between the mineralised clay of Abenab West, the Abenab breccia body and the other vanadate localities in the immediate vicinity is not yet clear. According to C. M. Schwellnus (1945) a red unconsolidated clay forms the major portion of the binding material of the Abenab breccia below a depth of 600 ft. Thus there is a possibility that this body is united in depth with the Abenab West zone. The vanadium mineralisation of the breccia bodies and that of the Abenab West clay appear to be analogous in that a small percentage of coarse, early

vanadinite was first deposited, followed and partly replaced by much more descloizite. This makes it not unlikely that the vanadium mineralisation in the whole Abenab area proceeded simultaneously and that the differences in crystallisation are merely due to variations in the depositional environment.

The migration of large amounts of zinc and lead from the oxidizing Abenab West deposit to the adjacent Abenab breccia pipe would have provided space for the extensive contamination of the former by foreign material.

The Supergene Period: Second Stage.

The change from semi-arid conditions to the present relatively humid climate of the Otavi Mountain region must have influenced the oxidation processes of the Abenab West ore body. It is believed that this change was marked by the cessation of the vanadium mineralisation, increased calcite deposition, the replacement of willemite by calcite and the alteration of still available sphalerite to its normal oxidation product smithsonite. The oxidation of galena to cerussite and anglesite appears to have proceeded uninterruptedly up to the present, and the formation of covellite and limonite probably continued as well. Pyrite was found deposited on smithsonite and must accordingly belong to the second stage of the supergene period, while a little quartz crystallised late in the paragenetic sequence. The position of malachite is somewhat uncertain.

Hydrozinkite is intimately associated with smithsonite and replaces stringers of the latter in sphalerite from drill core no. 23. As in the case of Leadville, Colorado (Loughlin, 1918), the formation of the basic carbonate may be ascribed to loss of free carbon dioxide by solutions locally impounded in cavities in the ore, and a resultant recrystallisation of smithsonite originally formed.

The supergene changes that took place during the humid stage are insignificant in comparison with those effected beforehand. This may be ascribed to any or all of the following factors: (1) the rather complete oxidation of the sulphides already achieved, leaving only protected remnants of galena behind; (2) excessive dilution of active solutions by the increased rainfall; (3) relatively short duration of the humid period up to the present. Available analyses of Abenab ground water indicate that the solution of the heavy metals has practically ceased: one part per million of lead is reported in Abenab borehole water but zinc was not determined. However, the high sulphate content (102 p.p.m.) is noteworthy; it may be due to the decomposition of disseminated pyrite in the country rock.

Weathering and erosion played its part in reducing the outcrop, until the mining activities of today brought the natural history of the ore body to a sudden end.

X THE ORIGIN OF THE VANADIUM

The genesis of various types of vanadium ore has presented a problem in many parts of the world. The prevalent opinion seems to be that nearly all economic concentrations of vanadium are secondary—the products of meteoric water action. In the case of the vanadates of the Otavi Mountain area, there are sufficient indications that their lead, zinc and copper content is derived from primary sulphides, but there is no agreement as to the source of the vanadium. The divergent views on this problem have been based mainly on field evidence, not on a detailed study of one or more deposits. The three principal genetic theories are summarised below and evaluated in the light of results obtained during the mineralogical investigation of the

Abenab West ore body. It is realised that any theory that may be proposed to account for the occurrence of vanadates in the Abenab West mine would be unsatisfactory unless it could be extended to explain the other deposits in the surrounding area as well.

General observations bearing on the genesis of the Otavi region vanadium deposits have been given by C. M. Schwellnus (1945) and are briefly as follows. The vanadates are either found intimately connected with the oxidized zone of sulphide deposits (never with the unaltered primary ore, e.g. Tsumeb, Berg Aukas, Abenab West) or as isolated occurrences whose areal distribution seems to be dependent on the vicinity of oxidizing sulphides. There is a correspondence between the zinc and copper content of such primary ore bodies and the composition of the vanadium minerals in the surrounding area. Crystalline crusts and concretions of vanadium ore, formed *in situ*, are found in sand-filled solution cavities extending for a few feet below the surface of the dolomite. Within these depressions descloizite has also been encountered in stalactitic and arborescent forms, suggesting cold water deposition. The larger purely vanadium deposits (Baltika, Uris, Abenab), occurring in fissure veins and breccia bodies with calcite, peter out within a few hundred feet from the surface, no matter whether they are situated high up in the mountains or low down in the valleys. Even the exceptionally large Abenab deposit became narrow and relatively poor at 700 ft., though it is still unknown to what depth the mineralisation extends.

These features have suggested to most authors a genetic relationship with lead-zinc-copper sulphides, a relatively recent near surface mineralisation post-dating the formation of cavities in the dolomite, and the agency of supergene waters during deposition. The first and obvious source to look to for the vanadium from which such secondary concentrations could be derived is the metallic deposits with which they seem to be more or less closely associated.

1. *Theory of supergene enrichment from sulphides.*

One possibility is that a primary vanadium mineral may be associated with the sulphides of lead, zinc and copper. H. Moritz (1933) attributed the presence of 0·005-0·01 % vanadium, spectrographically determined in certain specimens of galena from Tsumeb, to minute inclusions resembling patronite (VS_4) under the ore microscope. Patronite is only known from one locality, viz. Minasragra in Peru, where its association with hydrocarbons strongly suggests supergene conditions of deposition. Until corroborative evidence is forthcoming, the occurrence of patronite in hypogene sulphides must be regarded as very doubtful. The only other vanadium sulphide, also regarded as a possible hypogene source of vanadium (Ramdohr, 1950) is sulvanite, Cu_3VS_4, a very rare mineral not yet noted in ore deposits of the Otavi Mountain Land. The possibility that primary vanadium-bearing silicates may have been deposited with the sulphides must also be taken into account. This source has been held responsible, in part at least, for the descloizite and vanadinite of the San Antonio mine, Chihuahua, Mexico. (Hewitt, 1943.) No silicate gangue has been found in association with the sulphides of the Otavi Mountain region.

Another possibility is that vanadium may be present as a trace element in the common sulphides. A. Stahl (1926) states in favour of this theory that samples of "almost pure" galena from Tsumeb contained amounts of 0·034-0·041 % V_2O_5. G. E. Claussen (1934) established the presence of 0·0n-0·00n % V spectrographically in a few samples of pyrite and galena but not in sphalerite from various localities,

and this led W. H. Newhouse to ascribe the origin of most secondary vanadates mainly to this source. Other work has cast some doubt on the applicability of these results. A. C. Skerl (1934) was unable to detect vanadium spectrographically in galena and sphalerite from Broken Hill, Rhodesia, and A. W. Clark (1931) had previously reported the absence of vanadium in galena from Olifantsfontein and sphalerite from Berg Aukas. A suite of minerals from the Tsumeb mine was investigated spectrographically by H. Moritz (1933) who concluded that pure galena, primary chalcocite and quartz are devoid of vanadium whereas tennantite, enargite, sphalerite and the other sulphides as well as calcite and dolomite contain $\pm 0{\cdot}001\%$, with pyrite and secondary chalcocite somewhat more. From spectrographic determinations on galena from various sources B. Wasserstein (1945) concluded that this mineral cannot be the source of the vanadium, a view supported by V. M. Goldschmidt whose remark "galenite never contains vanadium" is quoted by Wasserstein. It seems that the only primary sulphide that may carry appreciable vanadium as a trace element is pyrite, which is only a subordinate constituent of sulphide deposits in the Otavi area.

At Abenab West samples of the massive sulphide vein encountered in drill hole no. 23 yielded the following assay values for vanadium:

0·1   0·05   3·30   0·06   0·10   % V_2O_5

The sulphides were not free from oxidation, however, and the high vanadium value was definitely found to be due to descloizite occurring as a secondary mineral encrusting limonitic material in a cavity. Mineragraphic study of polished sections from this drill core, of the disseminated sulphides in the country rock, and of the partly altered galena in the ore body itself, failed to reveal the presence of any primary vanadium mineral. Samples of pure galena, sphalerite and pyrite from these sources were submitted to Mr. W. J. Pienaar for qualitative spectrographic analysis, but no vanadium could be detected.

The primary sulphides at Abenab West seem to be analogous to the huge Tsumeb deposit and belongs to the same period of mineralisation. If vanadium was introduced with the sulphides, the largest vanadium deposit would also be expected near or at Tsumeb, which is not the case. It has been suggested that the Abenab West ore body might represent only the root of a zoned sulphide deposit, the copper-rich portion of which has been eroded. If this portion had contained appreciable vanadium, it might conceivably have been large enough to account for the large vanadium deposits in the Abenab area. However, in the Tsumeb deposit nearby there is no primary zoning (Willemse *et al.*, 1944), copper-rich minerals have been preserved, and the presence of vanadium in the primary ore has not been adequately demonstrated. If the theory of a primary zoning of vanadium is true, it is remarkable that not even traces of the metal are found in the lower levels of the ore body.

If vanadium, like arsenic, was derived from the primary sulphides, one would expect them together in a mineral like endlichite which is intermediate between vanadinite and mimetite. The fact that both mimetite and vanadinite are found in the Abenab West mine suggests that vanadium might have had another source.

Several authors (Clark, Skerl, Schwellnus) have argued that even under the most favourable circumstances the primary sulphides would be an inadequate source of the vanadium in the vanadates. Assuming that galena deposits which have since been eroded had carried as much as $0{\cdot}5\%$ V_2O_5, Schwellnus calculated that at least two million tons of galena would be necessary to account for a vanadium deposit of the size of Abenab.

The present investigation seems to confirm the idea that the primary sulphides may be eliminated as a source for the vanadium in the deposits under review.

2. *Theory of supergene concentration from country rock.*

The geological characteristics of the vanadium deposits, as summarised above, can be equally well explained if it is assumed that the lead, zinc and copper content of the vanadates is derived from sulphide deposits, while the vanadium was leached by ground water from the country rock. An external source for the vanadium is suggested by the peripheral relationship of the vanadates wherever they are associated with oxidized lead and zinc deposits (Schwellnus, 1945) though this feature would also be explicable in terms of supergene enrichment from the sulphides.

Vanadium is the tenth in abundance among the metals in the earth's crust and is markedly enriched in certain sedimentary rocks. This fact creates a better *prima facie* case in favour of the lateral secretion of vanadium than in the case of geochemically rare elements like lead or copper.

In support of the theory that the vanadium has been leached from the sediments of the Otavi System, Schwellnus cites the results of two chemical analyses, viz. 0·05% V_2O_5 in Otavi shale and 0·02% V_2O_5 in Otavi "stinkkalk". The first-named value is in agreement with several spectrographic determinations on the shale (Wasserstein, 1945) and is about twice the average value for shale. It is well known that shales in general show the highest and dolomite and limestone the lowest vanadium content among sedimentary rocks. Under the country rock theory, the shale horizons are regarded as the most likely source, though it is significant that most vanadate occurrences seem to be situated in calcareous rocks. The latter may have contributed vanadium in view of their relatively great solubility.

Objections that may be raised against this theory are the following:

(1) The shale was laid down in water and the vanadium did not pass off in solution. How could it subsequently be dissolved by ground water?

The answer probably lies partly in the composition and physico-chemical state of the ground water, and partly in the way vanadium is fixed in the shale. (See below.)

(2) Shale is an impervious rock and is resistant to ground water circulation. The calcareous rocks which are soluble, on the other hand, have a very low vanadium content.

(3) The volume of rock available to leaching waters is limited by the depth of weathering. Can this account for all the vanadium in the large deposits found?

The amount of shale containing 0·05% V_2O_5 that is required to give the 10,000 tons of V_2O_5 recovered from the Abenab mine (in round figures after Schwellnus, 1945) may be calculated as 20 million tons. This quantity would, for example, be contained in a portion of the third shale band at Abenab (stratigraphic thickness = 35 ft.) 26 miles along the strike and 50 ft. in the dip direction. This is not excessive in view of the fact that several other shale horizons are known in the vicinity, and numerous shaly layers occur in the platy limestone.

(4) Chemical analysis of water from different localities in the Otavi Mountains failed to reveal the presence of vanadium (Schwellnus, 1945, p. 67). Neither could it be detected spectrographically in water collected in the Abenab West mine by the author. It seems, therefore, that the solution and transportation of vanadium by ground water do not take place at the present day. What unknown conditions must have governed these operations in the past?

(5) Small quantities of sphalerite and galena are distributed throughout the country. If the vanadium is derived from the sedimentary rocks, lead-zinc vanadates should be much more common than they are. "Sand sacks" devoid of ore are more often encountered than those with ore.

To this it may be said that, under the supergene theory, two conditions must be fulfilled before the deposition of vanadates can take place: (1) vanadiferous shale must have yielded its vanadium and (2) the oxidation of the primary sulphides must have started to supply the metals in solution. Even today a large proportion of the sulphides encountered at the surface is perfectly fresh, while vanadiferous shale is restricted in occurrence.

Some of these objections against the country rock theory are further discussed below.

3. *Theory of derivation from Hydrothermal solutions.*

Little is known about the possibilities of migration of vanadium in hydrothermal solutions. The element is pronouncedly oxyphile and is not found with sulphides, as already emphasized. It does occur in Ni-Co-U-bearing veins formed at moderate temperatures, e.g. Joachimstal, Czechoslovakia. Roscoelite is a primary vanadium muscovite, rarely present in gold-quartz veins. A little mottramite has been found among the volcanic encrustations of Mt. Vesuvius. According to Zambonini and Carobbi (1927) it is not an autopneumatolytic product but was formed from alkaline vanadates in the emanations at a period after the solidification of the lava, through the agency of water.

During the crystallisation of magmas, however, vanadium seems to be segregates in the early stages, entering the crystal lattice of the ferromagnesian minerals and of magnetite and ilmenite. Magnetite and ilmenite are not very susceptible to weathering and therefore the only significant primary source of vanadium, according to existing knowledge, is the mafic minerals of basic rocks. It will be a welcome contribution to geology if it can be proved that appreciable vanadium has reached the hydrothermal stage, and would help to account for some economic vanadium deposits as well as the widespread occurrence of vanadium in sedimentary beds.

The hydrothermal theory of the origin of the vanadium deposits in the Otavi Mountain area has been somewhat inadequately presented in the literature. H. Korn and H. Martin, quoted by Schwellnus (1945, p. 62), state that in their opinion the vanadium mineralisation is connected with transverse disturbances related to the Cretaceous alkaline eruptives of South West Africa. However, evidence has been cited to show that the vanadates were localised by earlier sulphide deposits. It is probable that these, rather than the vanadates, were related to the Cretaceous movements.

The vanadates at Abenab West and elsewhere crystallised after most of the sulphides had been oxidized. Therefore any hydrothermal activity responsible for the introduction of the vanadium must have taken place long after the emplacement of the sulphides, and could hardly have belonged to the same epoch of mineralisation. There are three possibilities under the hydrothermal theory:

(1) Descloizite, mottramite and vanadinite, as found, are hypogene minerals directly deposited from epithermal solutions. Evidence suggesting the deposition of the vanadates by ascending solutions is their frequent occurrence below chert horizons, dipping at an angle of 45° in the Bobos-Karavatu area. The pipe-like breccia bodies of Abenab and Uitsab may perhaps be interpreted as volcanic vents. However, the former has been described as a cave collapse phenomenon (Schwellnus, 1945), and the origin of these remarkable features is still regarded as an open problem by the present author.

This is the least satisfactory theory because it explains neither the geologic relationship between sulphides and vanadates, nor the superficial occurrence of vanadates in "sand sacks". The fact that the vanadium occurs in its highest oxidation

state militates against the suggestion that it was directly deposited from hydrothermal solutions.

(2) The present distribution of the vanadates is due to the solution, transportation and subsequent reproduction of primary hydrothermal vanadium deposits, by ground water. This theory accounts for the "sand sack" occurrences, but the "primary deposits" are rather difficult to locate. Abenab breccia pipe descloizite showed no significant differences, on chemical and spectrographic analysis, from obviously supergene descloizite (pseudostalactitic descloizite collected by C. M. Schwellnus on the walls of a sand-filled hollow at Berg Aukas—see Table 3). Descloizite nowhere shows any sign of redeposition or corrosion. A specimen that might be interpreted as such was found in the Abenab West mine, but the smooth blunting and rounding of the crystals are much more easily ascribed to abrasion by an underground stream.

The possibility that vanadinite might be a primary vanadium mineral is suggested by pseudomorphs, casts and replacement remnants of an early, coarsely crystalline generation in the Abenab West mine. Large crystals of vanadinite, encrusted and partly replaced by descloizite, were also found in the old Abenab mine and at Okarundu. The author is not inclined to accept the idea that these were the source of the vanadium, for the following reasons:

(i) At Abenab West early vanadinite is not the only mineral showing signs of leaching and replacement—the others are mimetite, calcite and willemite.

(ii) Mimetite shows a perfectly parallel history to vanadinite and the early generation would likewise have to be regarded as primary. Neither mimetite nor vanadinite has ever been described from a hydrothermal assemblage or any other environment than the oxidized zone of an ore deposit.

(iii) In the Abenab area there are not sufficient signs of early vanadinite to account for all the subsequent vanadates.

This modification of the hydrothermal theory has no foundation until the "primary" vanadium deposits can be satisfactorily indicated.

(3) Hydrothermal solutions contributed vanadium to supergene waters containg lead, zinc and copper derived from oxidizing sulphides. This hypothesis explains all the geological characteristics of the vanadium deposits because it recognises the rôle of supergene agencies. It merely amounts to referring the vanadium to an unknown source if no other alternative is acceptable.

Additional evidence concerning the Origin of the Vanadium.

During the present research new evidence was sought on the conditions of formation of the vanadium minerals. The following considerations are thought to have a bearing on the genetic problem:

1. *Mineral associations*

The prevalent association of vanadates with coarsely crystalline *calcite* in the Otavi Mountain region is decidedly in favour of a supergene theory of deposition, for calcium bicarbonate is the dominant soluble constituent of ground water in a dolomite-limestone region. It is remarkable, however, that dolomite crystals are never found with the vanadates. Descloizite and calcite were commonly deposited in alternating fashion, the former invariably lining the walls of cavities. As pointed out by Fowler and Lyden (quoted by E. S. Bastin, 1939, p. 118) repetitions in the mineral sequence of an ore deposit may be expected under a meteoric hypothesis of genesis.

In view of the calcite-vanadate association, certain experiments by Notestein (1918) on the conditions of deposition of vanadates in Colorado and Utah are of interest. He claims that calcite readily precipitates vanadium from vanadyl sulphate solutions. This precipitate is soluble in solutions of alkali carbonates and bicarbonates. He concludes that vanadium may be leached by sulphate-bearing ground waters and precipitated by calcite.

In the Abenab West mine paragenetic studies clearly indicate that the vanadates are later than most of the secondary minerals. The closest associate of vanadinite and descloizite is limonite-goethite. The clay minerals kaolinite and montmorillonite are ordinary constituents of sedimentary clay and are probably derived from the weathering of felspar and from shale. Dickite or other hydrothermal clay minerals have not been found, though the possibility was suggested by the author to H. Heystek. The colourless tourmaline is possibly of hypogene origin, but it cannot be definitely correlated with the vanadium mineralisation and its origin is problematic.

The association of vanadates with a little wulfenite and the unusual secondary zinc mineral willemite at Abenab West, Berg Aukas and elsewhere is bound to have genetic significance. In the arid areas of the Western United States and Mexico small quantities of vanadinite, descloizite, willemite and more often wulfenite are characteristic of the oxidized lead and zinc deposits (Pough, 1941). In the oxidized zone of the lead deposits of Ksil-Espé in Kazakhstan vanadinite and earlier wulfenite likewise occur; here the origin of the molybdenum is referred to the primary ore and the vanadium to the country rock. (Yanishevsky, 1934.) As far as the mineral distribution is concerned, the Otavi Mountain district is mainly distinguished from these occurrences by the extraordinary abundance of vanadates in contrast with the rarity of molybdates. Under the supergene theory for vanadium, this difference must be explained in terms of the vanadium content of the country rocks and the molybdenum content of the sulphides. The frequent association of these rate minerals could hardly be fortuitous and must be ascribed to the operation of a common set of conditions. An adequate explanation of the peculiar mode of oxidation may be provided by the factor of climate. At any rate, there is little doubt that the willemite at Abenab West is a supergene mineral, which renders it unlikely that the vanadates were deposited by other supergene solutions.

2. *Zoning of the Abenab West ore body*

From the distribution of ore minerals in the Abenab West mine it is clear that vanadinite is restricted to the central part of the ore body, where it surrounds a lead-rich core; descloizite occupies a peripheral zone which overlaps on to the vanadinite zone, while willemite is found some distance away. These observations are best interpreted as the result of outward migration of zinc-bearing waters during oxidation of the original sulphides, while vanadium migrated from the outside inwards.

3. *Mineral Structure*

The vanadium minerals (especially descloizite) are characterised by a variable composition which is responsible for the occurrence of a series of mineral varieties differing only slightly in composition but markedly in colour and pleochroism. The zonal alternation of these varieties in descloizite crystals indicates a prolonged period of deposition, subjected to repeated variations in the composition of the depositing solutions. This deduction is in full agreement with a supergene theory of formation. The various pseudomorphs and replacement textures already described may likewise be explained by normal changes under supergene conditions.

4. *Temperature of crystallisation*

Minerals whose stability relations are known to provide points on a thermometric scale have not been encountered in the vanadium deposits of the Otavi Mountain Land. However, nearly all minerals contain primary microscopic liquid inclusions and these have recently become an important aid in geologic thermometry. The method consists of determining the temperature at which the contraction vacuole normally present in each inclusion disappears when heated. The temperature of filling is assumed to represent the temperature at which the liquid in the cavity was originally included. This temperature may be determined by using a microscope, heating stage and thermocouple. A new experimental procedure has been introduced by H. S. Scott (1948) by which the filling temperature is detected audibly instead of visually, by the explosive effects of the temperature-pressure relations in the inclusions upon the enclosing mineral. Under favourable circumstances, with the necessary apparatus and skill, this decrepitation method seems to provide a rapid and reliable means of estimating temperatures of deposition.

S. W. Bailey and E. N. Cameron (1941) have emphasized that the applicability of liquid inclusion thermometry is restricted to occurrences that satisfy all the assumptions on which the method is based. In general there seems to be agreement that the method is valid for minerals formed in open cavities at shallow depth, provided that a number of representative specimens are studied. The vanadium deposits of South West Africa seemed to be a suitable subject for a preliminary investigation along these lines.

The author is deeply indebted to Dr. F. G. Smith of the University of Toronto, Canada, who agreed to examine a series of minerals from the Abenab and Abenab West deposits by the decrepitation method which was developed in his laboratory. Eleven specimens which included the following, were prepared and submitted for the test: red vanadinite, red descloizite and greenish yellow descloizite from the Abenab breccia body (collected by Dr. J. W. Brandt several years ago); brownish drusy vanadinite, dark green descloizite, light brown descloizite, calcite, willemite and sphalerite from Abenab West; yellow vanadinite and green descloizite (replacing the vanadinite) from the Okarundu Prospect.

The following reply was received from Prof. Smith:

"I heated your minerals in the decrepitation apparatus, but could not detect any sounds which could positively be ascribed to filling of fluid inclusions. Each specimen decrepitates, but the rate curves are not of the liquid inclusion types. Calcite and dolomite have a very similar kind of decrepitation, which we are calling anomalous until we determine the cause. Soft crystals grown in water solutions characteristically give the anomalous decrepitation. The minerals deposited by surface waters give vigorous anomalous decrepitation, some of the softer hydrothermal minerals also give the anomalous effect as well as other stages, but metamorphic minerals rarely give it at all. The extremely vigorous anomalous decrepitation of your specimens suggests deposition from surface water. Absence of a liquid inclusion stage of decrepitation is not indicative of absence of fluid inclusions. If the minerals were formed by ground water, most of the inclusions would split the crystal at too low a temperature to be easily detected.

Under the microscope, fluid inclusions are difficult to resolve, but some were seen clearly in the vanadinite. They seem to be of one phase, probably liquid. If this is true for the others, it indicates that cold water deposited the minerals.

One experiment which I believe would aid in coming to a conclusion regarding the genesis of these minerals is an analysis of salts obtained by leaching the finely

ground minerals. (Grind to 20 mesh, clean, wash, grind with pure water in a clear pebble mill, filter, evaporate, analyse.) The salts obtained could be compared with the ground water salts and with published analyses of liquid inclusion water in hydrothermal minerals. The latter are usually rich in NaCl and low in $SO_4^=$.

It is unfortunate that the minerals did not give very positive results, but the indications are that they formed at low temperature in watery solutions, most probably in ground water."

According to a subsequent note received from Prof. Smith, the sphalerite and willemite specimens and presumably the two specimens from Okarundu as well, were of too small amount for good tests. The anomalous decrepitation of calcite is mentioned in a paper by Peach (1951).

An important conclusion to be drawn from these results is that the large red vanadinite crystals from Abenab showed the same decrepitation characteristics as the other specimens and may also be regarded as a low temperature product. This renders it more improbable than before that the analogous early vanadinite, partly replaced by descloizite at Okarundu and Abenab West, is of primary hydrothermal origin.

5. *Composition of Mineralising Solutions*

The only direct samples of the solutions from which the vanadium minerals were deposited are the fluid inclusions incorporated during crystallisation. At the suggestion of Dr. F. G. Smith attempts were made to determine the composition of these minute inclusions by leaching the finely ground minerals in water. 25-gram samples of Abenab vanadinite, Abenab descloizite and Abenab West descloizite were each crushed with distilled water in an agate mortar, filtered into a platinum dish and concentrated to a small volume. The cations were determined spectrographically after evaporating a few drops on purified graphite electrodes. For comparative purposes, Abenab West mine water and distilled water were treated in the same way. The following results were obtained:

	Principal Constituents	Appreciable	Little	Very Little
Abenab descloizite... ...	Si Mg Ca	Cu	B Na Mn Al	V Pb
Abenab vanadinite... ...	Si Mg	Ca Cu B V	Na Mn Al Pb	
Abenab West descloizite ...	Si Mg V	Ca Cu Pb Zn	B Na Mn Al Ni	
Distilled water				Si Mg Al
Abenab West mine water ...	Si Mg	Ca	B	Cu Mn Al

The variable amounts of Pb, Zn and V were not unexpected, as a suspension of very finely crushed material was seen to pass through the filter paper in each case and was removed by the centrifuge—though incompletely as the spectrograms show. The significant elements S, Mg, Ca, Cu, B, Na and Mn, however, show remarkable uniformity and with the exception of Na they are also present in the mine water in roughly the same proportions. Of course no comparison of the absolute amounts with those in the mine water would be possible, as the volume of the liquid inclusions is unknown.

The anions Cl^- and $SO_4^=$ could not be determined spectrographically and microchemical tests had to be employed. The only positive result obtained was for Cl^- in water leached from Abenab vanadinite, but these methods are evidently less

sensitive than the spectrograph; Ca, Cu, Pb and Zn were not detected either by microchemical methods.

The composition of the ore and gangue minerals themselves also affords an indication of the composition of the mineralising solution. That chlorine was a constituent, for instance, is proved by the presence of vanadinite and mimetite. Lindgren (1933, p. 861) explains the formation of cerargyrite during the oxidation of silver ores by "the universal presence of chlorine in waters, especially in those of arid climates." Combining the probably incomplete evidence from liquid inclusions and that from mineral compositions, it is clear that the solutions that deposited the vanadates contained the following elements at least: Pb, Zn, Cu, Fe, V, As, Ca, Si, Mg and Cl as major constituents and B, Mo, Na, Mn, Al, Sn, Ge, Be, Ni, Co, Ag as minor constituents and trace elements. (See Table 5.) It is obvious that most of the trace elements could have been derived from oxidizing sulphides together with Pb, Zn, Fe, Cu and As.

6. *Climate*

Climate has a profound influence on weathering and should, therefore, affect the possibility or likelihood of deriving the vanadium in mineralising solutions through this agency. Rankama and Sahama (1950, p. 599) state that solution and transportation of vanadium by weathering solutions take place especially in arid climates, while the ground water in a humid environment is particularly poor in that element. It has been shown that the Otavi Mountain region has formerly experienced relative aridity which is held to be responsible for the characteristic assemblage of secondary minerals including willemite, wulfenite, descloizite and vanadinite. It is now believed that the change of climate also provides an explanation, in terms of the supergene theory, of the fact that vanadium is not carried by ground water at present. In the opinion of the author, it is not necessary to invoke hypothetical ascendant waters for the solution of vanadium from the country rock.

The fact that many vanadium deposits in the Otavi Mountain area occur in solution cavities and karst phenomena in the dolomite implies either that these features could have formed in a relatively arid climate or that the arid period was preceded by a humid one. Schneiderhöhn (1920) who studied the karst topography of the Otavi Mountain area in detail, favours the former alternative and points out that karst forms depending on surface waters are lacking; moreover, there are forms that may equally well be explained by purely eolian agencies and by weathering through solution.

7. *Vanadium Content of Country Rock*

A series of qualitative spectrograms were made of rocks collected from various stratigraphic horizons between the second shale band and the grey laminated dolomite in the Abenab area. Only in the shale specimens could vanadium be detected, though it must be admitted that the presence of an excess of Ca in the others probably has a depressing effect on the intensity of the vanadium spectrum. Even thin shaly layers in the main laminated limestone showed appreciable vanadium while the surrounding limestone appeared to be barren. A small percentage of bitumen left after the solution of black limestone was tested but found to contain no vanadium.

The third shale horizon immediately below the Abenab West ore body has been weathered to a vertical depth of at least 340 ft. whereas the second shale band is fresh 200 ft. below the surface, where it is transected by the Malcolm Shaft. If the vanadium has been leached from shale by weathering solutions, the fresh rock may be expected to have a higher vanadium content than the weathered shale. To test this

assumption W. J. Pienaar kindly undertook quantitative spectrographic determinations on one fresh and two weathered shale samples. The method employed (Pienaar, 1951) is a modification of the method of Mitchell and Scott for the determination of the trace elements Mo, Co, V, Zn and Ni in soil and plant material. It essentially consists of concentrating the trace elements by precipitation from an acid solution with 8-hydroxyquinoline + tannic acid + thionalide and determining the amount of vanadium in the concentrate spectrographically by means of the cathode layer arc, using iron as an internal standard. The results are compared below with known figures for vanadium in shales.

Fresh shale from second shale band, Malcolm Shaft, Abenab West	0·024% V
Weathered shale from third shale band, 100-ft. level, Abenab West	0·025% V
Weathered shale from third shale band, 200-ft. level, Abenab West	0·026% V
Shale from Otavi system (Wasserstein, 1945) (·05% V_2O_5)	0·028% V
Average value for shale (Jost 1932, quoted in Rankama and Sahama, 1950) ...	0·013% V

If the results obtained are considered to be representative, they indicate that no leaching of vanadium has taken place. However, it is highly desirable that this conclusion should be verified by further analyses.

An interesting observation which has a bearing on the way in which vanadium occurs in the shale emerged from the spectrographic investigation. Preliminary efforts to extract the vanadium by crushing the shale in concentrated hydrochloric acid and even by boiling in aqua regia failed. The vanadium could be liberated only by fusion with sodium carbonate. The possibility that minute grains of vanadiferous magnetite or ilmenite may be the source of the vanadium was tested by making a heavy mineral separation, but with negative results. It seems, therefore, that the vanadium is intimately bound up in the shale and is only amenable to solution after total decomposition of the shale. According to Rankama and Sahama (1950) vanadium in shales is not detrital but occurs as the cation V^{5+} in clay minerals, where it neutralises the excess of negative charge caused by the presence of silicon. It remains there as long as the host mineral is unchanged. V. L. Bosazza (1940) has shown that vanadium in Australian and South African clays is not water-soluble until heated above 1000°C. No doubt the decomposition of clay minerals and the release of vanadium is ultimately effected by weathering also, especially in arid climates (Rankama and Sahama, 1950). The present study, however, seems to indicate that the weathering of the shale in the vicinity of the Abenab West mine has not yet reached that stage. It appears that the country rock theory must resort to higher lying material which has been totally decomposed and since removed by erosion, as a source of the vanadium in the ore. This is not too far-fetched, in the opinion of the author, because the eluvial deposits of vanadium ore prove that the vanadium deposits themselves must have been considerably denuded, and the shales no less.

Conclusion.

From the foregoing it should be clear that the balance of the evidence favours the theory that the vanadates were deposited by circulating ground water. The hydro-

thermal theory can only be maintained if the hydrothermal solutions are admitted to have lost their hydrothermal characteristics through cooling, contamination and extensive migration. No positive evidence of recent hydrothermal activity could be found in the Otavi Mountain region.

There are certain difficulties in ascribing the origin of the vanadium to the shale horizons of the country rock, but at present these still appear to be the most likely source. The following amendments are added to Schwellnus' theory of lateral secretion:

(*a*) The leaching took place during a relatively arid period when the ground water solutions were warmer, more concentrated and more acid than at present.

(*b*) The source of the vanadium was probably not the shale exposed at present, but higher-lying rock that was completely disintegrated and decomposed by erosion and weathering.

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Much of the information set out in these pages would not have been available without the co-operation of the following persons and institutions, to whom the author wishes to express his gratitude:

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PLATE I

A. Limonite rods, Abenab West mine.

B. Irregular limonite boxwork, Abenab West mine.

C. Weathered calcite with zonal structure, 340 ft. level, Abenab West mine.

D. Aggregate of white radiating willemite prisms, Abenab West Zinc Reef.

E. Willemite druse, Abenab West Zinc Reef. ($\times 2$)

F. Willemite rock, Abenab West Zinc Reef.

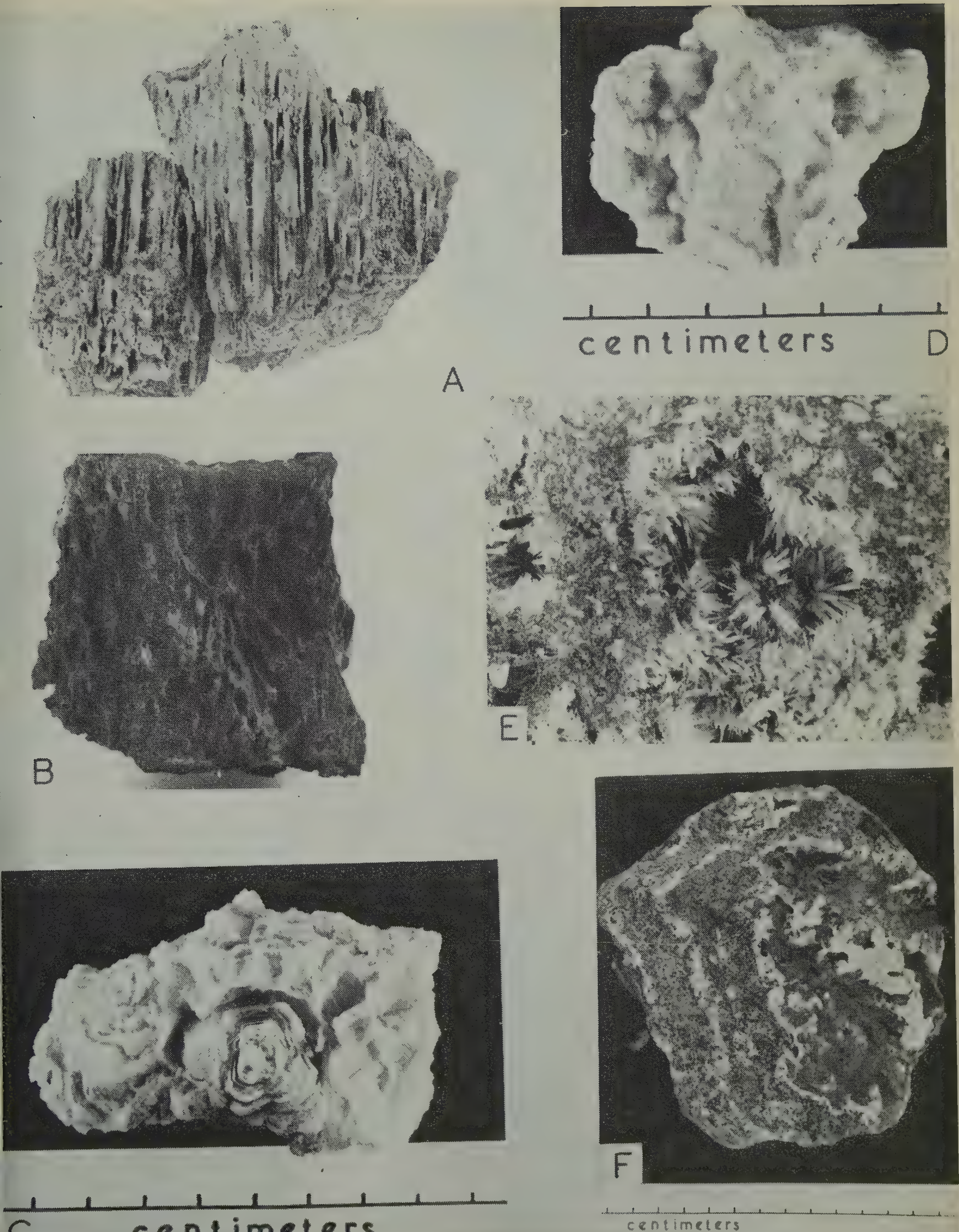


PLATE I

PLATE II

A. Greenish crystalline crust of descloizite, A. W. mine.

B. Greenish botryoidal crust of descloizite, A. W. mine.

C. Aggregate of descloizite crystals from W54, 340 ft. level, A. W. mine.

D. Abraded aggregate of descloizite crystals.

E. Thin section showing pleochroic zones in descloizite from W54, 340 ft. level. ($\times 4$)

F. Pseudostalactitic descloizite from "sand sack" deposit at Berg Aukas.
(Collected by C. M. Schwellnus).

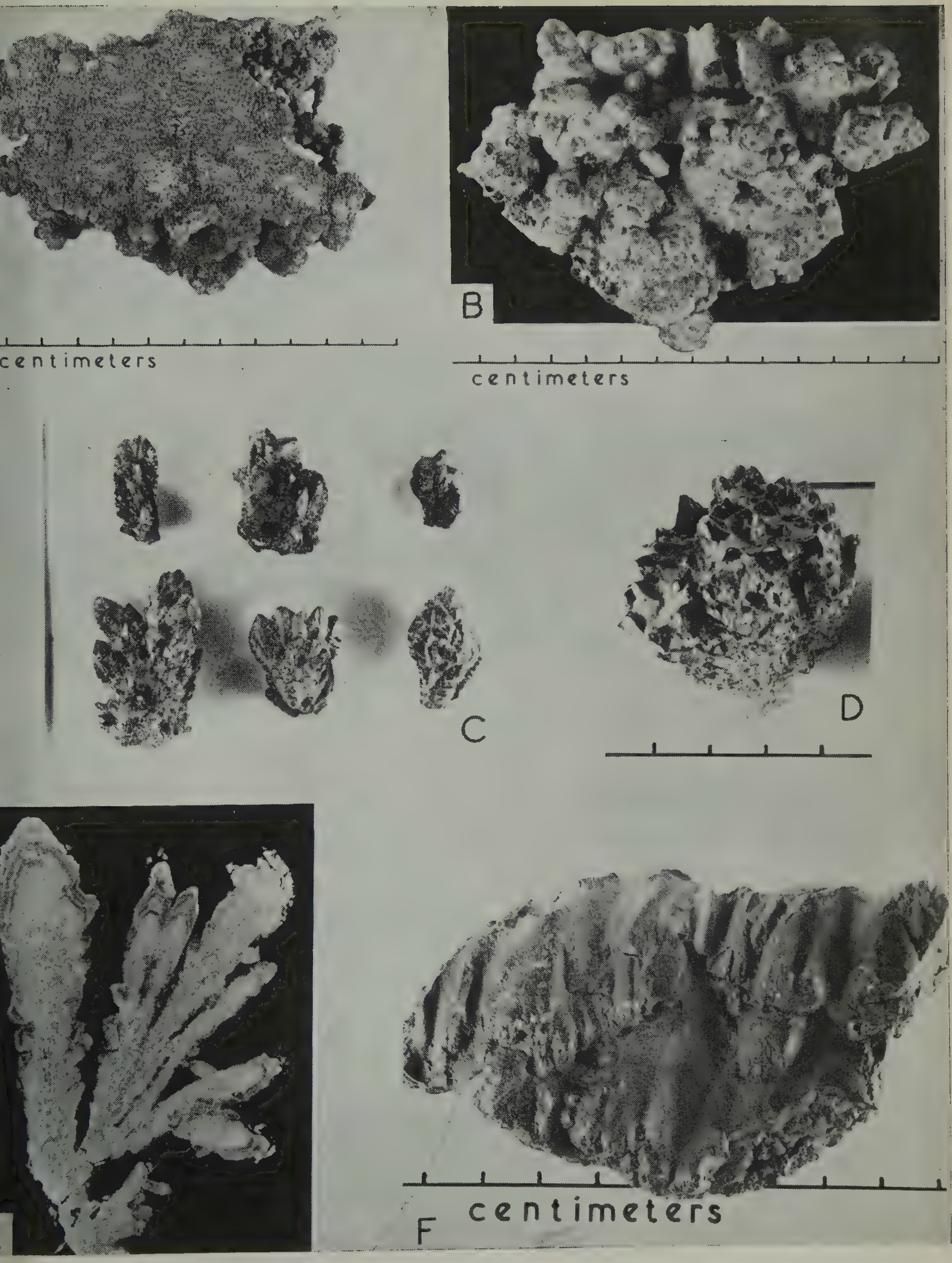


PLATE II

PLATE III

A. Radial aggregate of large red vanadinite prisms (Abenab).

B. Euhedral vanadinite crystals (Abenab).

C. Aggregate of minute straw yellow radiating vanadinite prisms (Abenab West).

D. Cuprian descloizite as hexagonal prisms pseudomorphic after vanadinite (Abenab West Zinc Reef).

E. Aggregate of relatively coarse vanadinite partly replaced by descloizite (340 ft. level, A. W. mine. See photomicrograph Plate IVA).

F. Yellow zonal vanadinite embedded in descloizite-calcite rock (Okarundu breccia-limestone).

G. Hexagonal pseudomorphs of descloizite after vanadinite, occasionally with light yellow remnants of the latter (Okarundu breccia-limestone).

PLATE III

PLATE IV

A. Descloizite (light) replacing vanadinite (dark), Abenab West mine, 340 ft. level. Crossed nicols. (×38)

B. Hexagonal shells of descloizite filled by calcite and pseudomorphic after vanadinite (Okarundu breccia-limestone). (×18)

C, D, E. Willemite party replaced by calcite, 340 ft. level, A. W. mine. (×156)

F. Red pleochroic zones in longitudinal section of descloizite, A. W. mine. Polarised light. (×32)

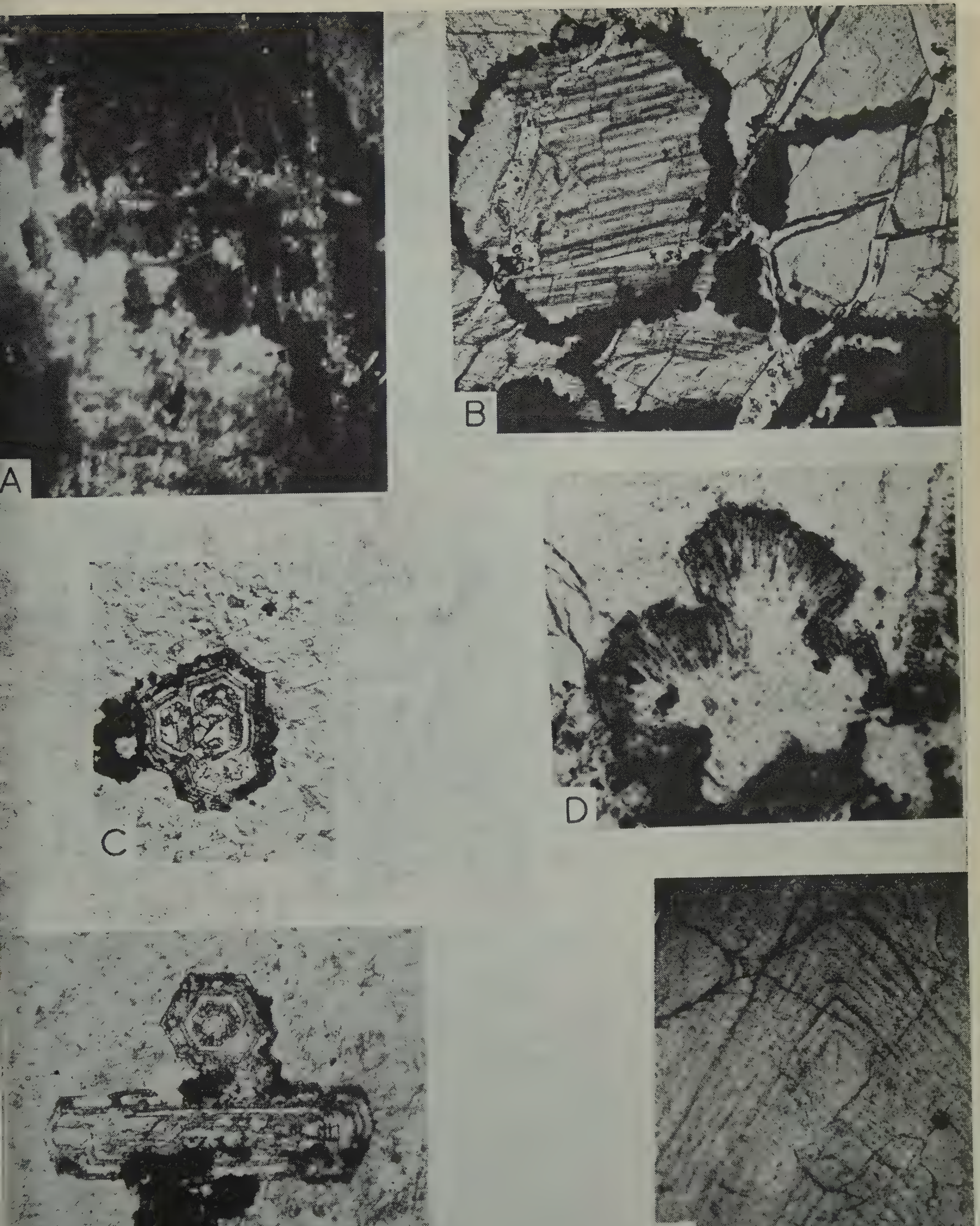


PLATE IV

PLATE V

A. Hydrozinkite surrounded by galena. Oxidised sulphide ore from drill core No. SDDH 23. ($\times$126)

B. Advance islands of willemite in a sphalerite bleb from the Zinc Reef. The black veins and rims consist of greenockite and the unidentified black mineral 5 (supergene galena ?) which are not clearly distinguishable in the photomicrograph. ($\times$70)

C. Zonal streaks of red-brown sphalerite in yellow-green sphalerite from a sulphide bleb in the Cluster Dolomite. ($\times$22)

D, E. Crystals of pyrite with characteristic mantles, disseminated in limestone country rock. Crossed nicols. ($\times$52)

F. Zonal calcite intergrown with red descloizite (small breccia body near the Zinc Reef, Abenab West). ($\times$4)

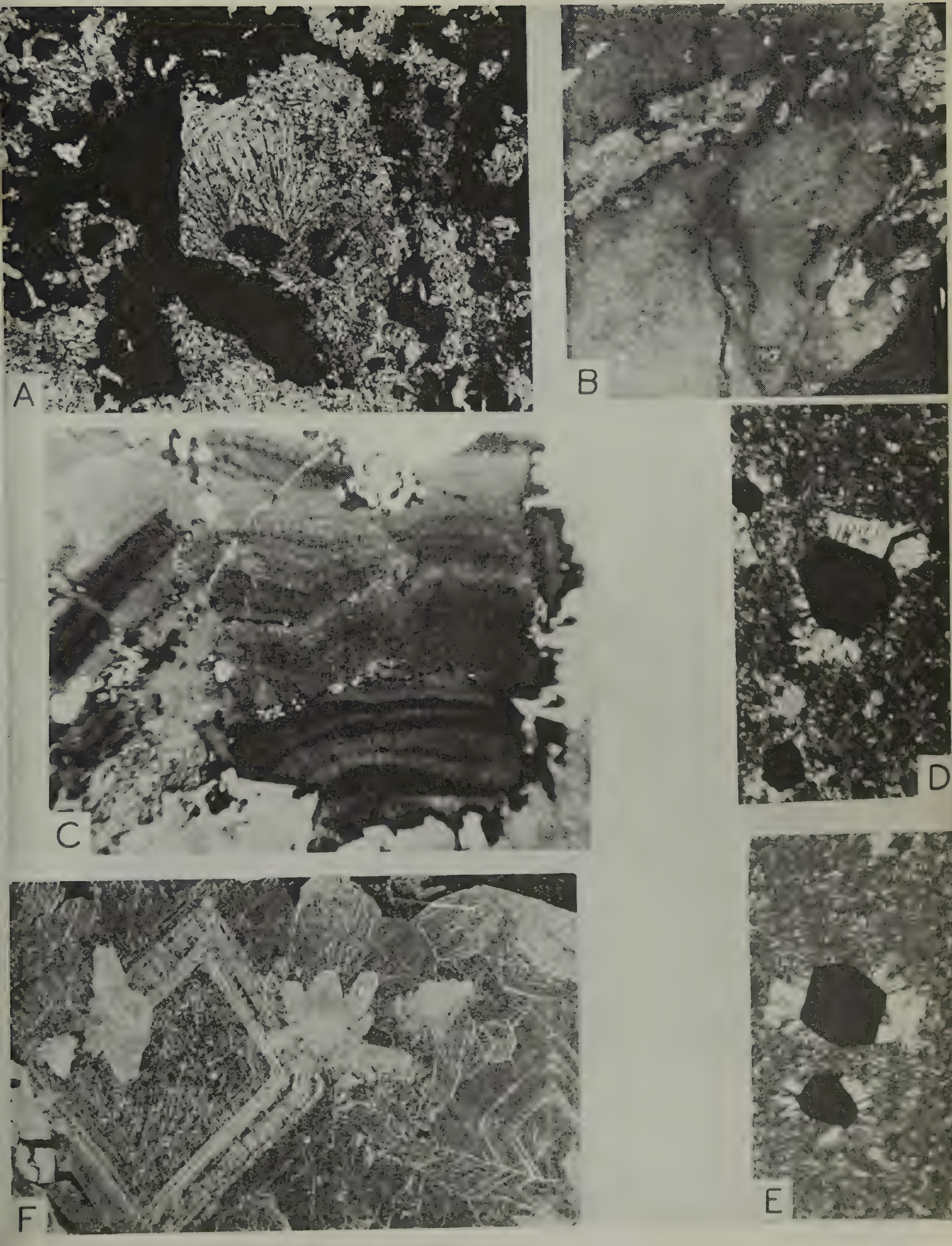


PLATE V.

THE APPLICATION OF THE "STEREOGRAPHIC PROJECTION" TO PROBLEMS IN STRUCTURAL GEOLOGY

By

V. VELLET, B.Sc.

(Submitted September, 1954)

ABSTRACT

Field and mining geologists encounter structural problems involving angular or linear relations. The customary methods adopted to solve these problems are either geometric or graphic in nature. Walter H. Bucher developed a new approach to such problems and showed the advantage of his cyclographic method. Jerome Fisher on the other hand exploited the usual stereographic projection as utilised in crystal analysis, and emphasised its advantage over the method proposed by Bucher in certain problems.

In this paper, 27 problems in field, structural and mining geology are presented. These are solved by utilising a method which is cyclographic as well as stereographic in nature. This aspect of the method is not new to crystallography, but represents a new approach to problems in structural geology.

This joint method which is three-dimensional in nature, is designated "stereographic projection" and involves only simple geometric concepts. Emphasis is laid on the distinct simplicity and other important advantages of the method.

This paper was awarded the Corstorphine Medal and the first prize of the Geological Society of South Africa for the year 1954.

CONTENTS

I HISTORICAL BACKGROUND

The concept of the stereographic projection first occurred in the writings of Hipparchus, who was the great predecessor of Ptolomy. Ptolomy's work on this subject appeared in Venice in 1558. From Ptolomy the knowledge of the projection passed to the Arabs and from them to mathematicians like Jordanus and Commandinus, in the 16th century. The application of this principle in the manufacture of astrolabes received much attention, in particular in the memorable work of Clavius in the beginning of the 17th century.

We owe our first impression of the application of the projection in solving spherical triangles to Adrianus Metias in his "Primum Mobile" (1633). This author is also credited with the first stereographic net. In the 18th century J. Harris, J. Wilson and C. Leadbetter used globes and spherical projections with much advantage in England. It would appear that the two fundamental properties of the projection, viz. angular truth and the fact that a circle on a sphere is projected either as a straight line or a circle, were common knowledge in the 18th century. Apparently the astronomer, Edmund Halley, published a paper in 1695, which directly influenced the development of the projection in the 18th century.

After the papers of W. Emerson (England) and G. S. Klügel (Germany), F. E. Neumann (1823) was the first to draw the attention of German scientists to the application of the projection in crystallography and he was followed by Miller in 1852. Since then the method has been extensively applied and developed, as is evident in the works of Federov (1893, 1903) and G. Wulff (1902). Both these authors introduced a stereonet of 20 cm. in diameter. It is, however, to the works of S. L. Penfield (1901) that we turn for the most comprehensive treatise on the stereographic projection and its application to the problems in crystallography. Since Penfield, the method has received universal attention in the study of crystals and rock-fabric analysis.

In 1920 Walter H. Bucher drew attention to the value of the stereographic projection in the study of joints. He was followed by O. Seitz and Rudolph Sokol (1927), who clearly pointed out its practical importance in applied geology. Masanoba Morishita utilised the method in solving fault problems in Japan (1937), and in 1938 and 1941 D. Jerome Fisher indicated how it could be applied to problems involving space relations. E. Ingerson (1942) showed the usefulness of the method in solving problems in lineation. W. Bucher in 1943 showed its application to diamond-drill hole problems. In 1944 J. Gilluly modified W. Bucher's method as proposed in 1943. It was, however, left to W. Bucher to find a wider application in structural geology in 1944. In 1946 K. Lowe utilised the method in solving linear problems in granite. In 1947 R. Beckwith showed how it could be applied to the solution of fault problems.

II GENERAL PRINCIPLES

In the past students of crystallography have found it a laborious task to solve the problems presented by crystals. The stereographic projection, however, has proved a handy tool in offering a graphic solution to their problems, as well as eliminating untold calculations.

The graphical solution of the problems presented by crystals is in the first place based on the principle of spherical projection. The construction of the latter is illustrated in Fig. 1. An isometric crystal containing the forms of a cube, octahedron

and dodecahedron is mounted in a sphere so that the centre of the sphere coincides with that of the crystal. The exact centres of the crystal faces are located and normals are drawn from these to intersect the circumference of the sphere. The points of intersection of the normals and the sphere locate the so-called poles of the faces.

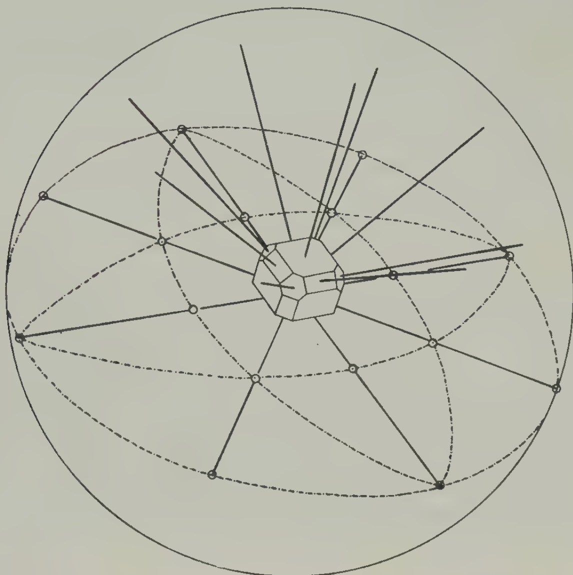


Figure 1

The construction illustrates the following important concepts:—

(*a*) Crystal faces which lie in one zone, or, alternatively, which are bounded by parallel edges have poles conforming to a great circle on the circumference of the sphere.
(*b*) All crystal faces which are parallel to the vertical axis of the crystal, or parallel to the vertical axis of the sphere, have poles which conform to the circumference of the equatorial plane of the sphere.
(*c*) If the plane-normals be extended, they will intersect in a point which is the centre of the crystal and the sphere.
(*d*) The angular distance between the poles of two faces, as measured on a great circle, is the same as the angle between the normals, as well as being the supplement of the solid angle between the crystal faces.

From Fig. 1 it is clear that the spherical projection is three-dimensional and thus still cumbersome to portray. For this reason the stereographic projection was developed.

The Stereographic Projection

The construction of this projection is also shown in Fig. 1. The face-poles, as located previously, are projected to the equatorial plane by connecting all poles in the upper hemisphere to the lower pole of the sphere and vice versa. Where these

connecting lines intersect the equatorial plane points are located which represent the (face-pole) stereographic projection of the crystal faces. In visualising this projection three-dimensionally, the point of vision would be the upper pole of the sphere. For convenience the upper hemisphere has been shown, as the conventional method involves the projection of the upper hemisphere only.

A most important feature of this projection is that all angular distances and directions which can be plotted only with difficulty on a spherical surface appear on the equatorial plane in such relations that they can easily be measured. The validity of this feature is indicated in Fig. 2 which represents a vertical section through a graduated sphere. AA^1 is the vertical axis and W-E the horizontal axis of the sphere. Let points A, B, C, and D be points of intersection between 0°, 20°, 40° and 60° arcs (on the sphere) with the plane of the section. Connect points A, B, C and D to A^1 and note where the connecting lines pierce the horizontal (o, b, c, d). Then draw radii to B, C and D respectively. Before proceeding any further it is desired to define an angle.

The circumference of any circle may be divided into 360 equal divisions. The circle in Fig. 2 represents such a circle. When it is constructed OA (the radius) is rotated around O as centre and will reach successive positions OB, OC and OD and so on until the position A is again reached—the circle then being complete. The amount of rotation from A to B is given by the arc-distance AB which coincides with 20 divisions of the whole (360). Thus AB subtends an angle of 20 degrees (1 division = 1 degree) at O the centre of the circle. Furthermore, when another circle is constructed inside the circle (Fig. 2), having the same centre O but radius OK, the radii OA, OB, OC and OD are intersected in K, L, M and N respectively. The arc distances KL, LM and MN are not the same as AB, BC and CD, but the angular

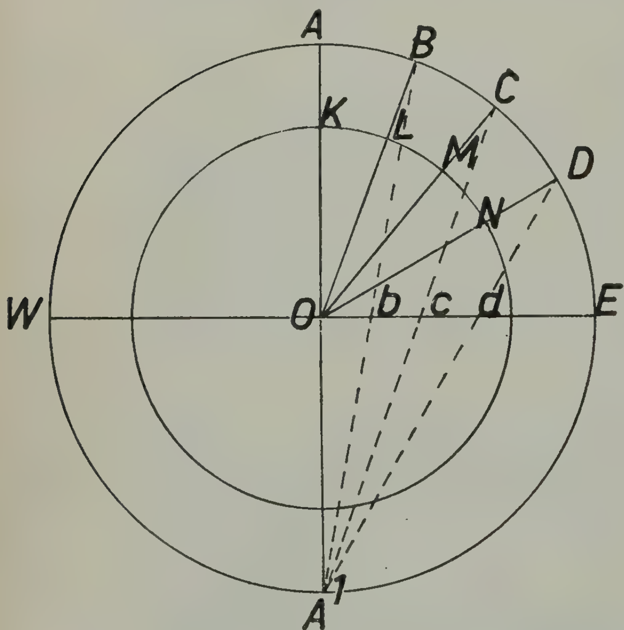


Figure 2

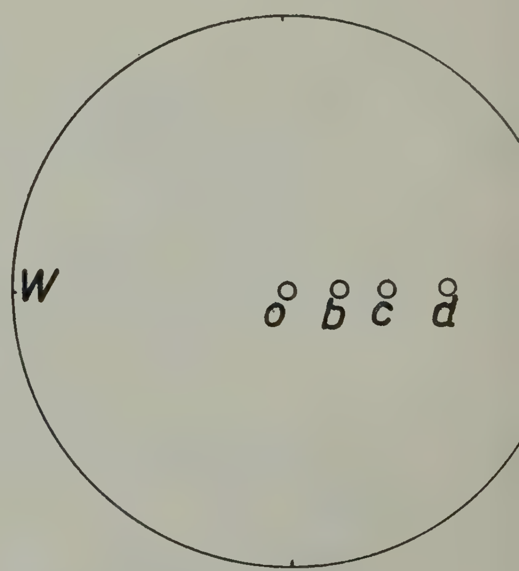


Figure 3

distance (20°) remains the same. Thus the arc distance will vary with the radius of the circle. On this principle the construction of the protractor is based.

As stated previously, the connecting lines between A, B, C and D and A^1 (lower pole) pierce the equator (WE) in O, b, c and d. The latter points represent the stereographic projection of crystal faces, of which AO, BO, CO and DO are the normals. It is thus important to realise that the angular distances ob, bc and cd also correspond to 20°, 40° and 60°. But the distances ob, bc and cd do not remain the same. This truth may be shown as follows:

According to a well-known theorem the angle AOB (20°) equals twice the angle AA^1B.

Thus in triangle OA^1b

$$ob = \tan 10° \times r = \cdot 1763r \text{ (r the radius of the circle)}$$

Similarly

$$oc = \tan 20° \times r = \cdot 3640r$$
$$od = \tan 30° \times r = \cdot 5774r$$

$$OE = \tan 45° \times r = r$$
$$\therefore ob = \cdot 1763r$$
$$bc = \cdot 1877r$$
$$cd = \cdot 2134r$$

It follows that there is a gradual increase in angular distance between the projected points, although the arc distance between the normals remains the same.

This illustration will suffice to show that the closer the points to be projected are to the upper pole, the closer will the projected positions group around the centre, and a gradual increase in angular distance will be evident as we move along the equator. For clarification the equatorial plane of Fig. 2 is reproduced in Fig. 3.

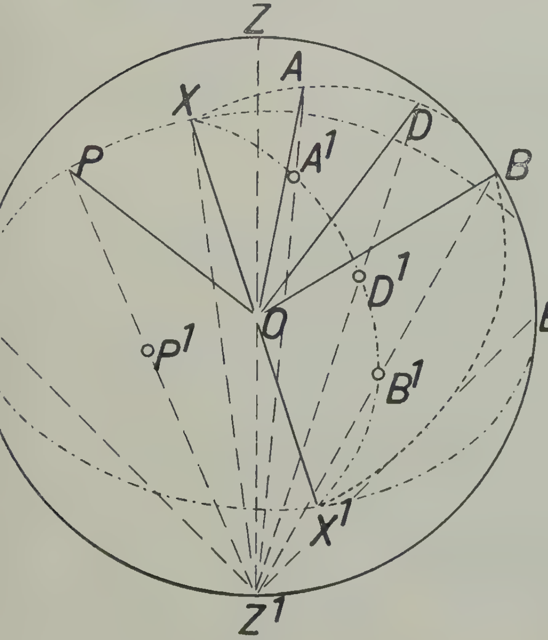


Figure 4

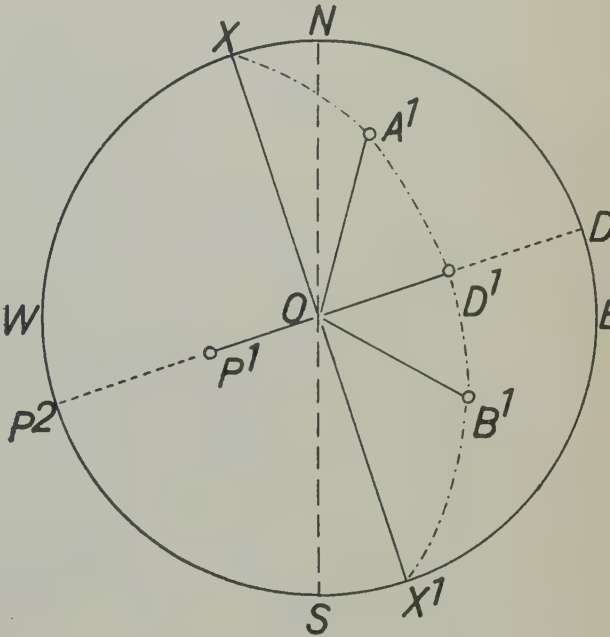


Figure 4(a)

The application of the projection is, however, not confined to the study of crystals. Any plane or line, if considered as passing through the centre of an imaginary sphere, may be projected in a similar way.

Fig. 4 represents a perspective view of a sphere with a vertical axis Z-Z^1, a horizontal axis W-E and a centre O. The upper half of a plane which pierces the equatorial plane of the sphere in XOX^1 intersects the circumference of the sphere along a great circle. AO and BO represent two arbitrary lines within the plane piercing the sphere in A and B respectively. Since the equatorial plane $WXEX^1$ represents the horizontal, XX^1 must be the strike of the plane under consideration. DO represents a line normal to the strike XX^1 and is thus the direction in which the plane dips. PO is a line vertically disposed with respect to the plane XOX^1, and it pierces the sphere in P. Lines are drawn from points in the upper hemisphere to the lower pole (Z^1) of the sphere and the points of intersection with the equatorial plane noted. Thus the stereographic projection of points A, D, B and P conform to A^1, D^1, B^1 and P^1. The upper pole (Z) projects at O (the centre); W, E, X and X^1 remaining the same as these points are confined to the equatorial plane.

Fig. 4(*a*) is a bird's-eye view of the equatorial plane and thus serves as a stereogram showing the stereographic projection (P^1) of the plane as well as the cyclographic projection of the plane ($XA^1D^1B^1X^1$). P^1O represents the trace of a line normal to the plane—the angular distance P^1D^1 being 90°. The angular distance P^1P^2 is the amount (in degrees) of inclination of the plane-normal; similarly the angular distance D^1D^2 gives the angle of dip (from the horizontal NESW) of the plane. NS is the trace of a vertical plane which has been ideally chosen to represent a geographic north-south direction. Thus the line P^2O (or P^2D^2) gives the direction in which the plane-normal is inclined. It is measured on the circumference of the equatorial plane as the angular deviation ND^2. Similarly the strike of the plane (XX^1) is measured by reference to N, the north pole of the equatorial plane.

Figs. 4 and 4(*a*) illustrate the principles involved in the application of the projection to problems in structural geology as used in this paper. This method which involves the projection of the plane-normal as well as the plane, will for the sake of convenience be known as the "stereographic projection".

Bucher's Projection (Cyclographic Projection)

Walter H. Bucher, who has contributed appreciably to the application of the stereographic projection to structural geology, has developed another method of projection, which has found much favour with students who are less acquainted with the usual method.

The principles involved are indicated in Fig. 5. The lower half of a sphere is projected any desired distance to a plane which is parallel to the equatorial plane. O is the centre of a sphere and lies on AB, which is a meridian and also represents the strike of any desired plane dipping in the direction OC. Thus the arc ACB may be regarded as the intersection between the plane and the sphere. Lines are drawn from P (the imaginary upper pole of the sphere) to the circumference of the equatorial plane and are extended the desired distance towards the plane of projection.

The equatorial plane of the sphere projects as a circle; thus points A and B will automatically project on to the circumference of the circle at points A^1 and B^1. The centre of the sphere (O) projects as O^1, the centre of the circle. Fig. 5 is a perspective view of the whole operation and Fig. 5(*a*) shows the projection in plan.

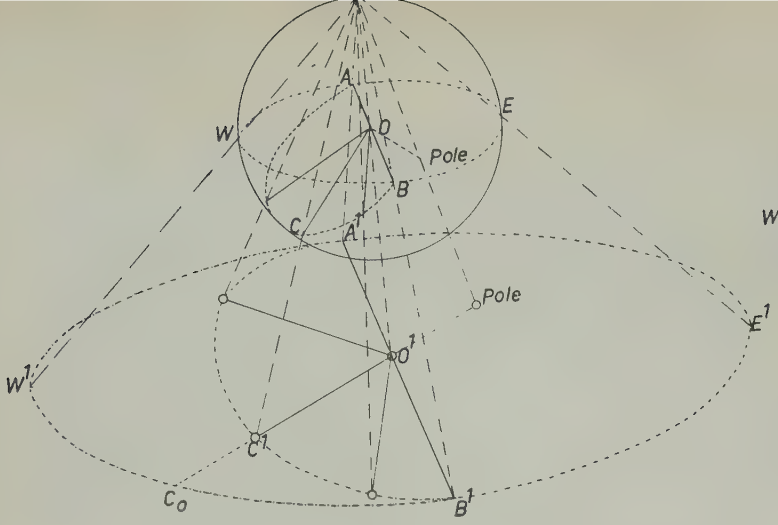


Figure 5

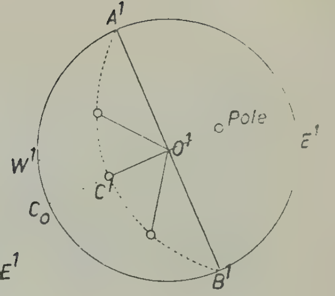


Figure 5(a)

It will be noticed that the plane ACB projects as $A^1C^1B^1$; O^1C^1 being the direction of dip. The angle between ABC and the equatorial plane of the sphere is given as C^1C_0 (W. H. Bucher, 1944.)

III. THE WULFF NET

It has been shown that a two-dimensional projection of a crystal can readily be utilised to find the angular relation between crystal faces. To measure angles a stereonet has been designed and for our purpose a Wulff Net will be adopted. The construction of the net may be visualised as follows:

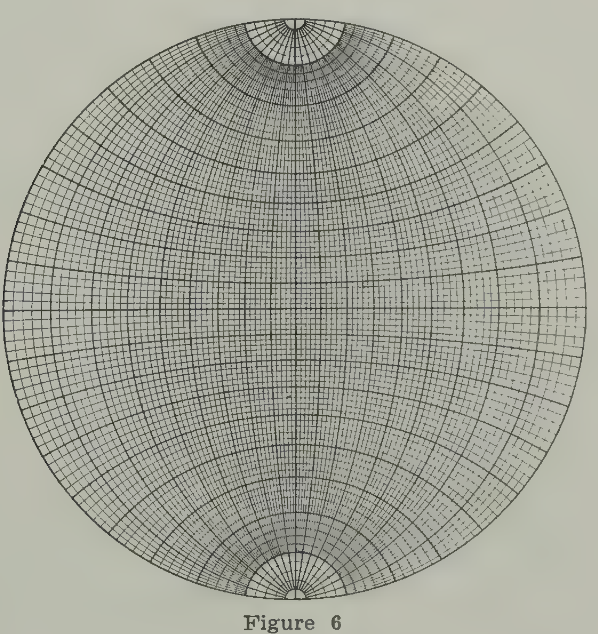


Figure 6

Regard a globe, with lines of longitude and latitude, orientated such that its north-south axis is horizontal and that the south pole points towards the observer. In this way the latitudinal or small circles as well as the great circle of the equatorial plane will be perpendicular to the line of vision. Regard the meridional great circles as intersections of planes (passing through the north-south axis) and the surface of the sphere. These planes will therefore, all have a common trend parallel to the line of vision. Now the portions of these planes lying in the upper hemisphere are cyclographically projected on to the horizontal plane containing the north-south axis. They will project as great circular arcs all terminating in two common points (the north and south poles) on the circumference of the plane of projection. The vertical plane containing the north-south axis of the globe will project as a straight line through the centre of the plane of projection. This line terminates in the same two common points mentioned above, and it constitutes the north-south axis of the plane of projection. The great circle of the equatorial plane of the globe projects as a straight east-west line and is referred to as the equator of the plane of projection. The latitudinal small circles when projected, form small circular arcs on either side of the equator. When referring to the Wulff net the terms small and great circles will, for convenience, be used instead of small and great circular arcs.

The circular plane of projection resulting from the cyclographic projection of a globe as indicated above is a Wulff net (Fig. 6). Great and small circles appearing on the net are 2° apart, and every tenth-degree line is accentuated so as to catch the eye and facilitate readings. A thumbtack is pierced through the centre of the net which is usually glued to a rectangular piece of cardboard, and a square piece of transparent paper placed upon it, so that the paper rotates freely around the centre. When data is transferred to the paper the latter is known as a stereogram.

The function of the net may be illustrated as follows: Any arbitrary point on the stereogram signifies the point of intersection between a line (in space) through the centre of the sphere and the circumference of the sphere. The angle of plunge of that line, regardless of its orientation, is usually ascertained on the equator (but also on the N-S axis) of the net. This truth is realised if it is remembered that the great circles, as they appear on the sphere, may be regarded as arcs of intersection between the circumference of the sphere and planes striking in a north-south direction. Thus the function of the net is to represent the orientation of structural elements in such a way that their interrelations are easily determined.

IV. THE PRACTICAL IMPORTANCE OF THESE PROJECTIONS

When solving the angular relations between crystal faces on a stereogram the initial step is the stereographic projection of all the face poles, the positions of which are determined by the interfacial angles obtained by goniometric measurement. But once this is attained the solution of the problem involves points, great-circles and small-circles. Thus lines as well as planes form an integral part of the solution.

The study of structural geology entails the angular relations between planes, and planes and lines in space. The geologist is familiar with the graphic and geo-

metric solution of problems pertaining to foliation, lineation, apparent and true dips, bedding and thickness of strata, fractures, joints, veins, faults, fold axes, cleavage, and flowliness of intrusives. He has perhaps found it no easy task to locate desired positions for diamond-drill holes and sometimes the implication of core-bed and core-foliation angles may have escaped his notice. Perhaps the time involved in graphic solutions has invoked a distaste for such problems, and the latter are solved by drilling a few more holes.

The three methods of projection, as outlined above, are ideally suited to solve any problem involving angular relations. The advantage of these methods both in time and efficiency, as well as in offering a solution on a small piece of paper, rarely with the aid of formulae or tables, overshadows all standard methods. However, as with all methods, each of these has its own specific limitations.

When the usual stereographic projection is used without the aid of planes, as suggested by Fisher (W. H. Bucher, 1944), a concise comprehension of mathematics and perspective is presupposed. On the other hand the stereogram is usually a "point-diagram" which may become extremely involved, and the tendency would be to solve the problem purely mechanically. But it has its advantages when dealing with steep planes or when transferring a plane to any desired position by rotation. These advantages will be clarified in due course.

"Bucher's Projection" is ideally adapted as an aid for those field geologists who are not familiar with the usual stereographic method of the crystallographer. One disadvantage is that the lower hemisphere is projected on to the equatorial plane. Anyone who adopts the method is required to visualise a plane projecting below the surface of the stereogram.

The writer has had the opportunity of using all three methods and has found the cyclographic projection distinctly more practical; moreover, while lecturing in structural geology to students who fortunately had a knowledge of the stereographic projection, he observed that a combination of the two methods found favour. On the other hand when only the usual stereographic projection was applied, it became clear that to locate the plane-normal, the plane proper had to be plotted in any case. Thus the joint data could be exploited with ease.

In an effort to find a suitable name for a method which is neither wholly cyclographic nor stereographic, it was decided to employ the term "Stereographic Projection." In an abbreviated form it will be known as the S.P. method. Whenever we refer to "Bucher's method" or the usual stereographic method, these will be known as the B.P. and P.P. methods respectively.

In the following pages various problems in economic and structural geology are solved by using the S.P. method. It is impossible to cover a wide field, and instead type-problems are presented in order to serve as a guide and stimulation within the scope of this fascinating field.

Some of these problems have already appeared in literature. In this paper solutions to these problems were obtained by using a somewhat different approach.

V THE PROBLEMS

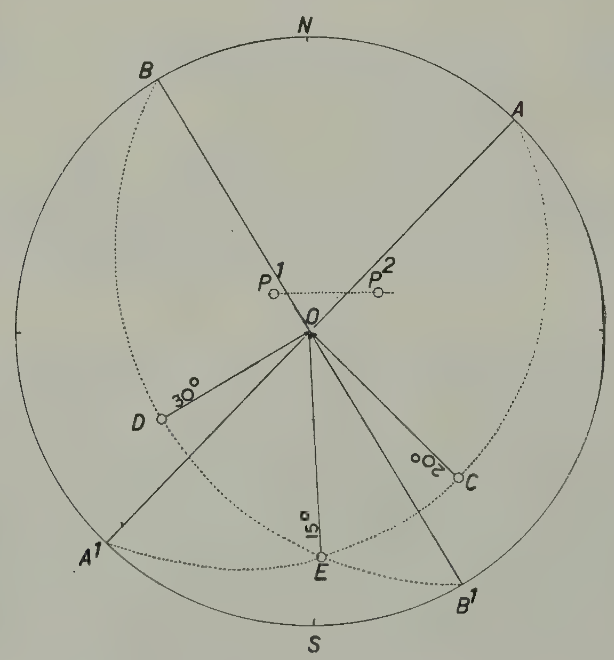
Problem 1

Two planes AA^1 and BB^1 have the following attitudes:

	Strike	*Dip*	*Direction of Dip*
AA^1	N45°E	20°	N45°W
BB^1	N30°W	30°	N60°E

Question: Determine the trend and plunge of the line of intersection.
Procedure: Place a square piece of tracing paper (having the same size as the net) over the stereonet, carefully piercing the paper at the centre with the thumbtack which is fixed to the former. The paper must rotate freely. While facing the net locate the geographic north and south positions and indicate on the stereogram as N and S. Through the centre of the net, plot the lines AA^1 and BB^1 which represent the strike of the two planes.

To portray the attitude of the planes in space, proceed as follows: Since we are using the S.P. method, our point of vision is the upper hemisphere in the spherical projection. Thus a plane dipping east will pierce the sphere in the westerly upper portion and its projection will naturally intersect the equator, somewhere between the centre of the net and the periphery. Similarly a horizontal plane when projected coincides with the circumference of the net, and a vertical plane is represented by a line through the centre of the net terminating on the periphery. Thus the amount of inclination of an inclined plane is measured off from the periphery of the net.



Stereogram 1

To plot the plane ACA^1, the tracing paper is rotated until AA^1 coincides with north-south line on the net, and 20° is measured off from the right along the equator, this position being indicated by point C. To facilitate the interpretation of the stereogram, draw a line CO (O being the centre) making a small arrow at O indicating the direction in which the plane dips. Indicate the amount of dip in degrees (20°) on line CO, which is the dip vector. Similarly, plot the plane BDB^1; the dip of 30° being indicated on DO. Rotate the tracing paper until BO and DO respectively coincide with the equator and draw in the great circles through C and D.

It will be noticed that the curves ACA^1 and BDB^1 intersect in a point E. At the same time note that EO represents a line common to both planes. Thus EO must represent the line of intersection between the two planes.

Rotate the tracing paper until point E coincides with the equator (always use the easterly half of the equator as a reference line). It will be noticed that point E is located 15° from the periphery, thus indicating a plunge of 15° in the direction EO.

To measure the direction EO, extend the line to the circumference of the net, using the equator as reference line. Now note the angular deviation from the geographic north position along the periphery of the net.

Answer: By reference we find the line of intersection to plunge 15° in a direction N2°W.

Alternative method: We may also adopt the P.P. method.

Procedure: Follow all the steps as shown above but refrain from recognising EO as the line of intersection. Rotate the tracing paper until C coincides with the equator. While holding the paper in this position, count off 90° from C to the left, marking this position as P^1. P^1O represents a line which is normal to the plane ACA^1. The position P^1 is known as the stereographic projection of plane ACA^1; we may refer to it as the pole of ACA^1. Proceed similarly to establish the pole of plane BDB^1 (P^2). Whereas, two lines in space through a mutual point (in this case O) establish the attitude of a plane, a great circle through P^1 and P^2 would represent a plane containing both normals. This plane will dip 75° in a direction S2°E. The normal to the latter must be the line of intersection between the planes ACA^1 and BDB^1. By reference this normal coincides with EO and plunges 15° in the direction N2°W.

The above problem is most enlightening, as it offers clear proof of the following concepts:

(*a*) Any great circle (stereogram) through the pole of a plane will intersect the plane in such a way that the angle between the pole and the point of intersection is always 90°.

(*b*) If the trend of two lines be known, a plane may be constructed which includes both lines.

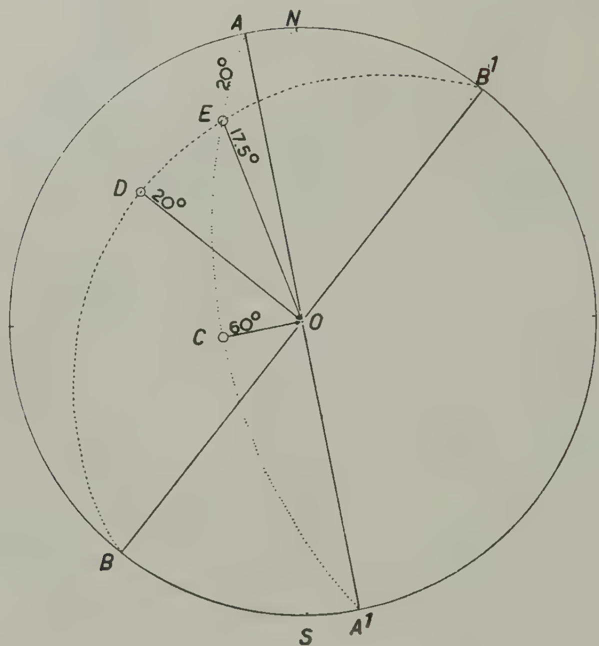
(*c*) If a plane be known, the plunge as well as the pitch of any line within that plane may be readily determined.

Problem 2

The average strike and dip of the flanks of a fold are given as:

	Strike	*Dip*	*Direction of Dip*
AA^1	N10°W	60°	N80°E
BB^1	N40°E	20°	S50°E

Determine: (*a*) The plunge of the axis of the fold.
(*b*) The pitch of the axis of the fold as recorded on both flanks.



Stereogram 2

Procedure: Consider the flanks of the fold as planes and plot the strike and dip, following the same procedure as outlined under the S.P. method. It will be noticed that in this case the two planes dip in the same general direction, whereas problem 1 treated a case where the dips were in opposite directions.

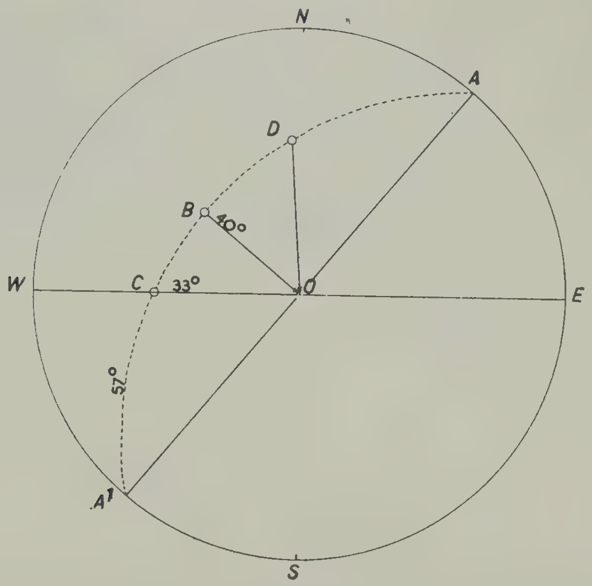
(*a*) The two planes having been plotted (ACA^1, BDB^1) locate EO the line of intersection, which establishes the axis of the fold. It is seen that the axis of the fold plunges $17\frac{1}{2}°$ in a direction S20°E.

(*b*) According to definition, the angle of pitch of the axis of the fold is measured from the horizontal within the plane in which it occurs. The strike lines AA^1 and BB^1, as recorded on the stereogram, represent the horizontal. Thus the angles EOA and EOB^1 are the respective angles of pitch of EO within planes ACA^1 and BDB^1. These angles are measured on the appropriate great circles of the dip vectors CO and DO by rotating the paper until each of these vectors coincides in turn with the equator. The pitch of the axis of the fold is 20° as given on the flank AA^1, and 62° as given on the flank BB^1.

Problem 3

Pitch is not a common element in the terminology of folding, but is more important when we think in terms of lineation. The ease with which problems pertaining to linear elements may be solved by adopting the S.P. method, is clearly shown in the following simple example.

Recorded strike of a bed is given as N40°E and the dip as 40° S50°E.
Determine: The apparent dip of the bed in an easterly direction.



Stereogram 3

Procedure: Plot the bed as the plane ABA[1] in the usual way. Draw a line through the centre of the net representing the projection of an east-west perpendicular plane. It will be noticed that the latter intersects the plane of the bed in C.

According to definition the angle of apparent dip is measured in a vertical plane as the angle of plunge of the line of intersection between the vertical plane and the plane of the bed. It follows that the apparent dip in an easterly direction is 33°.

There are a multitude of apparent dips between the zero position at A and the true dip at B. Similarly DO is also a direction of apparent dip. A linear trend confined to a plane is ordinary lineation, thus the problem serves as a basis for the following deduction. If AA[1] and BO should represent the strike and dip of a foliation plane, and CO a lineation caused by a linear arrangement of rod-like minerals, then the lineation will plunge 33° and pitch 57°.

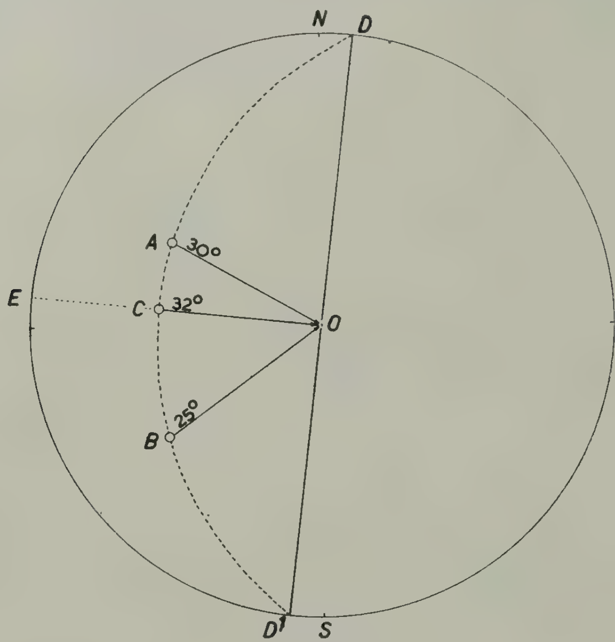
Problem 4

Many geologists, particularly when employed on underground mapping, have encountered the problem of measuring the dip of a bed when cuts are poor or tunnel-

faces dirty. It will be shown here that the desired true dip may very quickly be established by simply recording two apparent dips.

The following readings were taken:

Directions of apparent dip	N55°E;	S60°E
Amount of apparent dip	25° ;	30°



Stereogram 4

Procedure: Plot the two apparent dips, representing them as the two linear elements AO (30°) and BO (25°). These two lines must fall within the confines of the bed and thus establish the attitude of that bed in space. It is possible to reconstruct the latter by making the two lines coincide with the same great circle. Rotate the tracing paper until the points A and B coincide with the same great circle on the net. Draw in this great circle, noting where it cuts the equator (C) and the periphery of the net (DD^1).

Now DD^1 represents the strike of the bed and if CO is held on the equator the angle EC records the angle of dip.

Answer: Strike of bed N7°E; dip 32° S83°E.

Problem 4 (a)

In certain mining areas, drifting parallel to the strike of the beds is deemed more economical because some ore deposits are elongated parallel to the strike. In this case the strike of a bed is very easily measured, but owing to the absence of "markers" the dip may be elusive. In this case bedding may appear to be horizontal because of parallel bedding streaks on the walls of the tunnel. If a linear element should develop in the roof of the tunnel, the geologist is satisfied that the bed dips

towards the side wall. In this case small protuberences on the side wall may display an apparent dip.

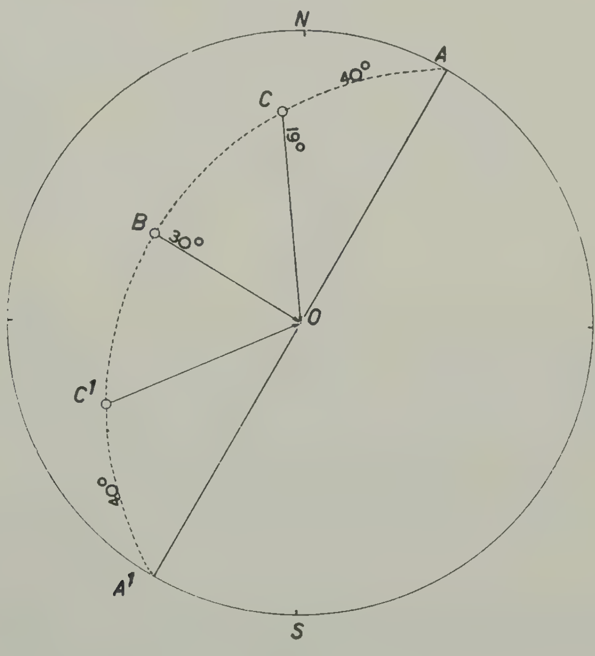
Say the strike was DD^1 as in problem 4, and the apparent dip was AO. Rotate the stereogram until DD^1 coincides with the north-south meridian (of the net), draw the great circle through A, and note where it intersects the equator (C). The dip of the bed has been established (EC), as in the former case.

Problem 5

In certain mining areas the distribution of ore bodies may be related to lineation, and the importance of recording the latter is far-reaching. Sometimes, however, it is extremely difficult to measure the space components of the lineation and for that reason perhaps many geologists refrain from recording it. Fortunately, it is easy to record the pitch of the lineation and if the S.P. method is adopted all the information is quickly determined (W. H. Bucher, 1920)

Data: The foliation and dip of a gneiss and the angle of pitch of rod-like minerals. Foliation N30°E, Dip 30° S60°E, Pitch of lineation 40°.

Determine: The trend and plunge of the lineation.



Stereogram 5

Procedure: Plot the foliation as the plane ABA^1; BO being the dip vector (30°). Orientate the stereogram with BO on the equator and locate point C on the great circle ABA^1, and 40° from A. The angle AOC now represents the angle of pitch. The line CO represents the lineation. Rotate the paper until C coincides with the equator

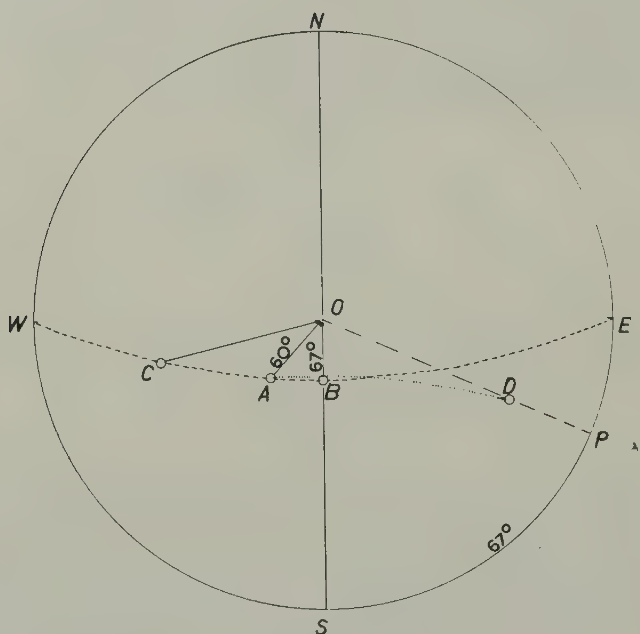
and read off the angle of plunge (19°). By reference the direction CO is given as S6°E.

It will be noticed that C^1O also satisfies the requirements of a pitch of 40°. However, it is easy to note in the field in what general direction the lineation pitches, and as there are only two possibilities, the desired answer is easily deduced.

Problem 6

A vertical N-S section is prepared through a mine and, for the sake of convenience, taken through the collar of a diamond-drill hole. The latter plunges 60° in a direction N42°E.

Question: What would be the angle of inclination of the diamond-drill hole (D.D.H.) as appearing on the N-S section?



Stereogram 6

Procedure: On the tracing paper plot the diamond-drill hole as the dip vector AO, plunging 60° in a direction N42°E. Draw a line N-S connecting the north and south pole of the stereogram to represent the trace of a vertical plane within which we wish to observe the D.D.H. Rotate the tracing paper until the line NS is on the equator, S coinciding with the geographic position west on the net. Draw a great circle through A, intersecting the equator in B and terminating on the edge of the stereogram in W and E respectively.

BO is the desired projection of the D.D.H. on a vertical N-S section, and the plunge of 67° is readily recognised.

The mechanics involved is comprehended if it be realised that AO may be regarded as an apparent dip of an E-W plane which dips north.

Another D.D.H. plunging 30° in a direction CO would give the same angle of dip when projected on to a N-S plane. Thus the stereogram offers a visual proof of the practical limits of projections when compiling sections.

Alternative method:

It would facilitate matters appreciably if we could look down on the flat of the N-S plane. To accomplish this it is required to rotate the plane through 90°. With the trace (NS) of the vertical plane occupying the north-south position of the net, count off 90° from O (the centre) to the right. O reaches the position E and the plane is now horizontal. The latter covers the whole field of the net, but let us consider only the semi-circle NES.

During rotation the diamond-drill hole must be transferred also to a new position in space. Thus A must shift through 90° to the right, reaching position D. The direction in which it has moved is indicated by a dotted line on the small circle AD. Now looking down on the plane of the N-S section, we can see the transferred D.D.H., DO plunging towards the centre of the net. Rotate the stereogram until DO coincides with the equator; draw in this line and extend it to cut the circumference in P.

Now PO represents the trace of the D.D.H. as appearing on the horizontal N-S section. Bearing in mind that NS is now a horizontal line it becomes obvious that the angle POS gives the angle of dip of the D.D.H. on the N-S section. By reference angle POS = 67°.

By holding NS in the north-south position of the net, check the latter result by restoring the plane to a vertical position (rotation through 90° to the left). During the process position P reaches position B, thus BO is the trace of the D.D.H., when the direction of observation is parallel to the N-S plane.

Problem 7

In a certain area a vertical D.D.H. intersected gold ore intimately associated with schistose amphibolite. The angle between the schistosity and the axis of the core was recorded as 40°. Owing to the complete absence of outcrops an open-cut was commenced in the most likely position. The face of the cut was inclined at 34° in a direction N55°E and intersected the amphibolite. The latter appeared as lines of intersection (CC Fig. 7) on this face and the pitch was recorded as 55°.

Question: What is the direction of dip of the schistosity.

Procedure: Plot the face of the "cut" as the plane ADA^1, DO being the dip vector (34°). With AA^1 coinciding with the north-south position of the net, count 55° from A along the great circle AD, locating position C. CO represents the line of intersection between the schistosity and the face of the cut.

The vertical D.D.H. recorded the complement of the angle of true dip. Thus the schistosity must dip (90° — 40°), which equals 50°. Rotate the stereogram until position C coincides with a great circle having a dip vector of 50°. Draw in this great circle intersecting the circumference of the net in BB^1.

BB^1 represents the strike of the schistosity. But it becomes obvious that the great circle B^2B^3 also satisfies the requirements of a 50° dip for the schistosity.

Thus there are two possible answers to the problem viz. the direction of dip of the schistosity is either N31°E or S20°E.

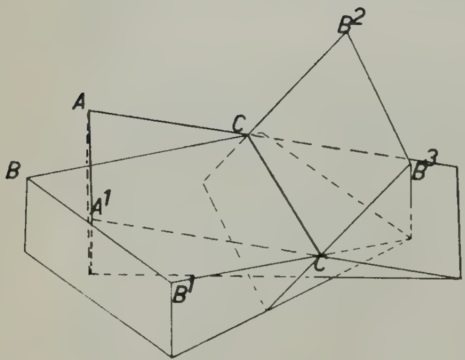
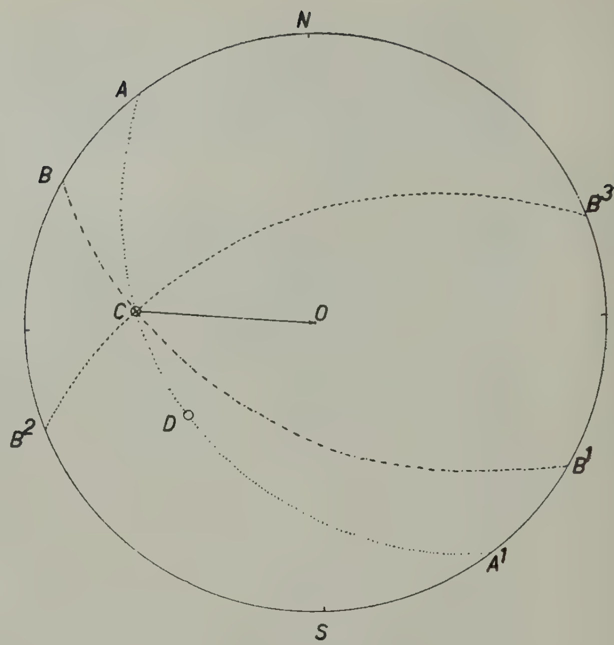


Figure 7



Stereogram 7

Problem 8

Three vertical diamond-drill holes A, B and C, spaced as shown in Fig. 8 with collar elevations reduced to a mean level, intersect a key bed at 200′, 100′ and 500′ respectively.

Question: What is the dip and strike of the bed.

In practice there are three standard graphical solutions to this problem, but all these methods entail either memorising an equation or representing a three-dimensional problem on a plane surface. It will be shown that the S.P. method has the distinct advantage of eliminating these procedures.

Procedure: A glance at Fig. 8 shows that the bed must dip towards AC. At the same time we note that the directions BA and BC are directions of apparent dip, the angles of dip being easily deduced from the tan-formula.

$$\tan \alpha = \frac{100}{300} \qquad \therefore \quad \alpha = 18° \ 30'$$

$$\tan \beta = \frac{400}{400} \qquad \therefore \quad \beta = 45°$$

Transfer the directions BA and BC either directly (by superimposing tracing paper on map) or by measurement on to the tracing paper. The respective directions are given as S55°W and S25°E. Plot the apparent dips as the vectors A^1O (18° 30′) and C^1O (45°). Rotate the stereogram until A^1 and C^1 are on the same great circle and note where the great circle cuts the equator.

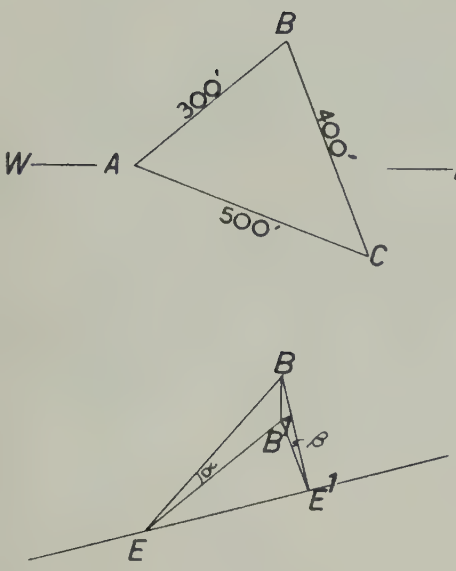
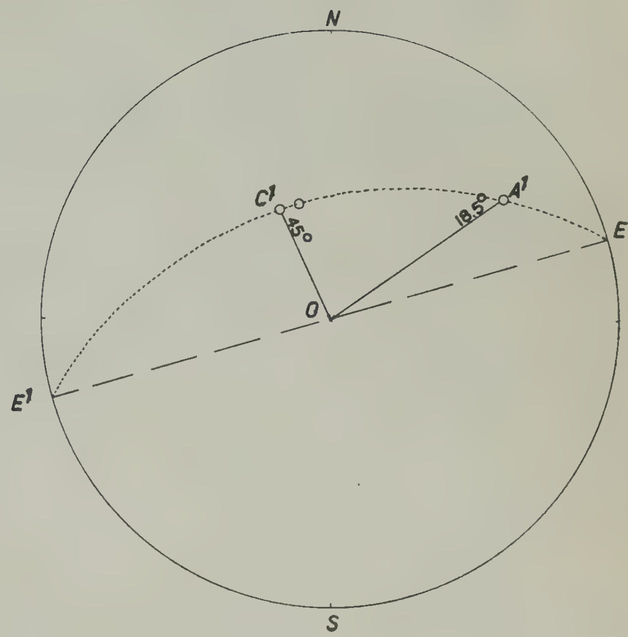


Figure 8



Stereogram 8

By reference, the strike of the bed EE^1 is given as N47½°E, and the dip as 45½° in a direction S15½°E.

It will be noticed that the direction BC (Fig. 8) very nearly coincides with the direction of true dip.

Problem 9

A conglomerate bed as exposed in a stope displays a dip of 64° in a direction S46°W. The long axes of the individual pebbles plunge south (Fig. 9).
Question: If it is assumed that the pebbles were not deformed during folding, what would be the arrangement of the individual pebbles when restoring them to the original plane of deposition.

See Fig. 9 and Stereogram 9 on following page

Procedure: On tracing paper plot the dip vector DO of the bed and establish the strike line CC^1 normal to it. Draw in the trace of the conglomerate bed CDC^1. Rotate the tracing paper until the N-S positions coincide with the north-south direction of the net. Note where this direction, i.e. the direction of plunge of the pebbles, truncates the bed (A). Now AO represents the individual pebbles which plunge south.

By reference, record the angle of plunge (55°) and the pitch (66°) of the pebbles.

In order to visualise the original attitude of the long axis of the pebbles, we must restore the conglomerate bed to a horizontal position. With DO on the equator rotate through 64° around CC^1 as axis. The position D transfers to D^1 and the bed is

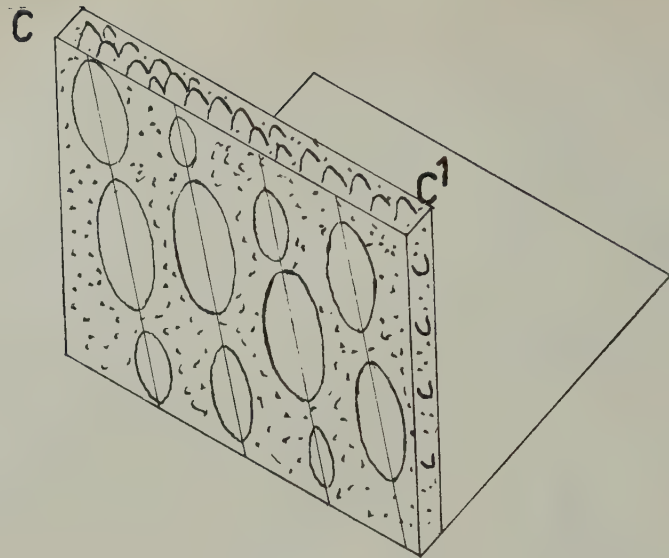
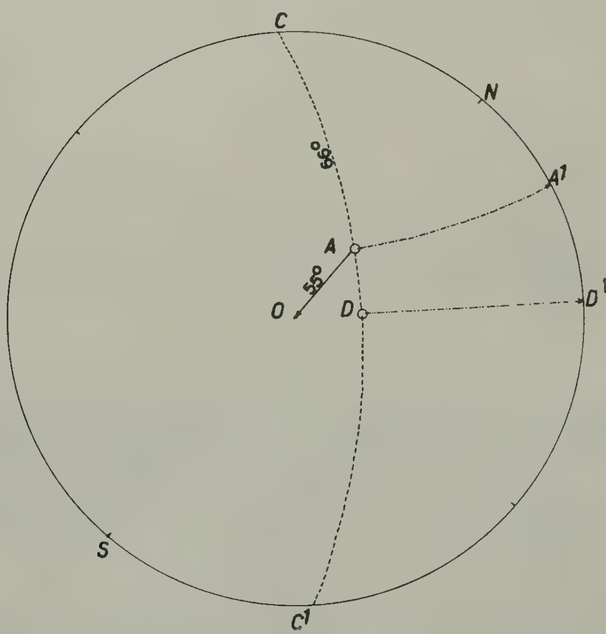


Figure 9



Stereogram 9

now horizontal. Similarly any position on the circumference of the great circle CDC^1 will transfer along small circles to new positions on the circumference of the net. Thus position A transfers to A^1 as indicated. The direction A^1O must be the original orientation of the long axis of the pebbles. This direction is given as S22°W.

It will be noticed that this original direction makes an angle of 66° with CC^1 (the strike of the bed). This angle is the same as the angle of pitch of the long axis of

the pebbles. Thus the method has a twofold result. It not only solves the problem efficiently, but also shows that the angle of pitch (if not available) may be determined easily.

A notable feature of the Rand gold deposit is the association between the long axis of pebbles and the distribution of pay-streaks. Thus the above problem could serve as an illustration of the adaptability of the S.P. method in problems involving projections of ore-shoots for exploratory purposes (L. Reinecke, 1927).

Problem 10

An unconformity separates two series of rock strata. The beds below the unconformity strike N45°E and dip 32°NW. The beds above the unconformity dip 75° east (Fig. 10).

It is obvious that the bed below the unconformity suffered two phases of disturbance.

Question: Reconstruct the lowermost beds, when the beds above the unconformity reposed in a horizontal position.

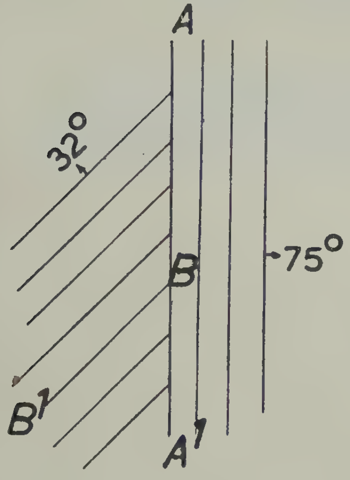
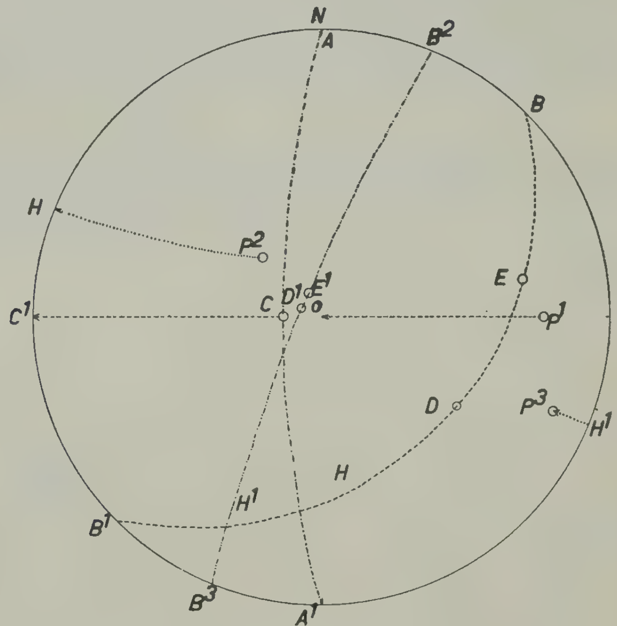


Figure 10



Stereogram 10

Procedure: Transfer the strike of the beds below and above the unconformity, either by measurement or superimposition on to the tracing paper. The beds above and below the unconformity are respectively given as AA^1 and BB^1. Plot the respective dip vectors of the beds as CO (75°) and DO (32°). Now the two planes ACA^1 and BDB^1 represent the two components on both sides of the unconformity.

Locate the respective poles of the two planes by proceeding as follows: Rotate the stereogram until C and D coincide in turn with the equator, and count off 90°

L

either to the left or right. These manoeuvres locate positions P^1 and P^2; P^1O and P^2O are the respective plane-normals of the beds ACA^1 and BDB^1.

It is desired to restore the beds above the unconformity to a horizontal position. Proceed as follows: With position P^1 on the equator rotate through 75° to the left around axis AA^1. P^1 is now in the centre of the net, the plane-normal being vertical and the beds (ACA^1) above the unconformity horizontal. Similarly P^2 is transferred to a new position P^3, the angle P^2P^3 measured along a small circle being 75°. It will be noticed that the latter rotation is accomplished in two steps. First rotate through 66° reaching the circumference of the net in H. Rotate the stereogram until H coincides with the equator and locate position H^1 on the directly opposite side of the equator. Rotate back to original position (P^1 on the equator) and count off 9° along the small circle H^1P^3, thus locating P^3. The total count is 75°.

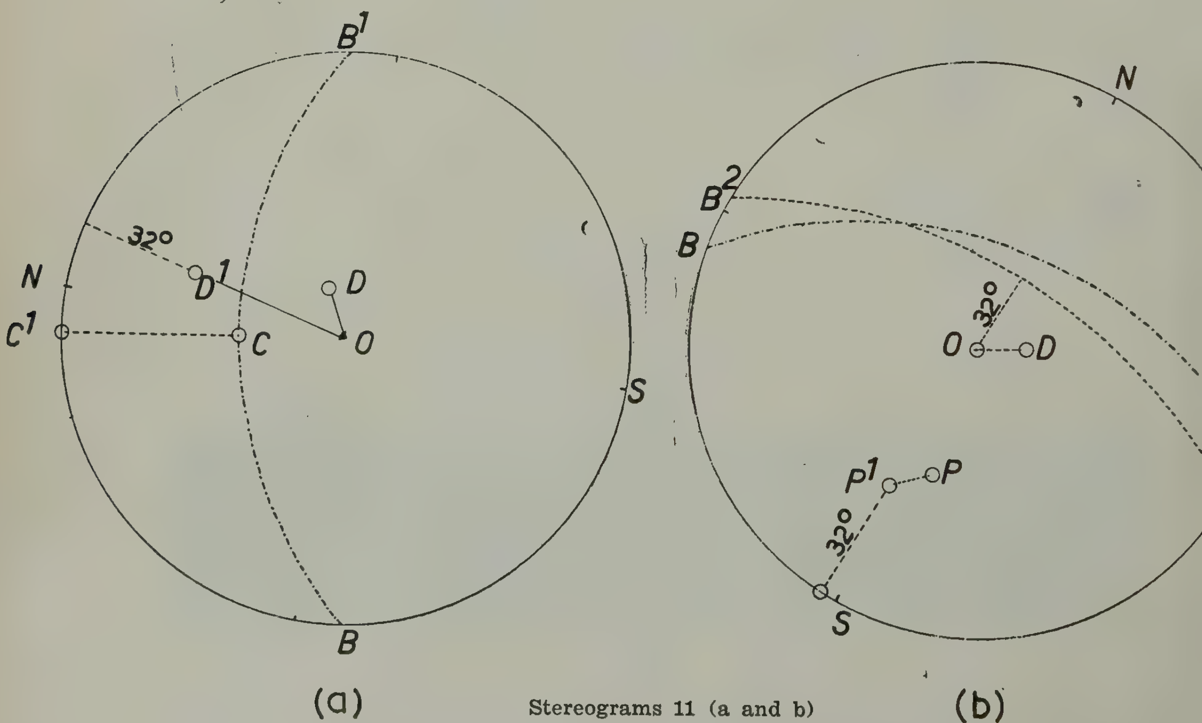
The position P^3 is the pole, so P^3 is the plane-normal of the reconstructed beds below the unconformity. With P^3 on the equator locate the plane $B^2D^1B^3$, which represents the original attitude of the beds below the unconformity before disturbance of the beds above the latter.

By reference it is shown that the original strike was N22½°E and the dip 81½° in a direction S67½°E.

This problem is very similar to one used by Walter Bucher in his paper. But, as previously stated, the S.P. method is different to the method adopted by him as reference to his paper will clearly show (D. J. Fisher, 1938).

Problem 11

An important key-bed dips consistently at an angle of 50° in a direction S10°E. A diamond-drill hole inclined 70° in a direction S60°W intersects this bed.
Question: On examining the core, what would be the angle of inclination of the bed as measured against the axis of the core. (Note that we are dealing with unorientated core.)



Stereograms 11 (a and b)

Procedure: (*a*) On tracing paper plot the bed as the plane BCB^1; BB^1 being the strike normal to the direction of dip CO (S10°E). Also plot the D.D.H. as the line DO.

If an inclined D.D.H. intersects a horizontal bed, the core-bed angle registers the true angle of plunge of the bore-hole. Thus it would facilitate the solution of the problem if the bed BCB^1 were transferred to a horizontal position. This is accomplished by rotation.

With CO on the equator, rotate through 50° around BB^1 as axis. Position C transfers to C^1 on the circumference of the net. Similarly position D transfers along a small circle through an angular distance of 50° to reach position D^1. Now D^1O represents the D.D.H. as it would appear in space if the bed were horizontal.

Rotate the stereogram until D^1O coincides with the equator and note the angle of plunge (32°). Thus the core-bed angle as required in the problem is 32°.

(*b*) The problem can also be solved by rotation around an axis normal to the D.D.H. This method is based on the concept that a vertical D.D.H. intersecting an inclined bed registers the complement of the true angle of dip.

Proceed as follows: Plot the pole P of the bed in the usual way. With DO on the equator rotate through 20° around an axis perpendicular to DO. D has transferred to the centre of the net, the diamond-drill hole now being vertical. Similarly P transfers to P^1. Rotate the stereogram until P^1 coincides with the equator and locate the plane of which P^1O is the normal. This reconstructed bed dips 58° in the direction of OP^1.

The trace of the plane of this bed (B^2B^3) is located 32° away from the centre of the net. The angle between this plane and the D.D.H., which is now in the centre of the net, is readily determined as 32°. In accordance with the definition the core-bed angle must be 32° (stereogram 11(b)).

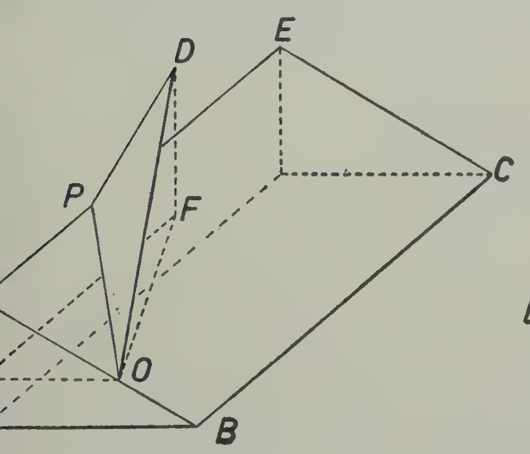
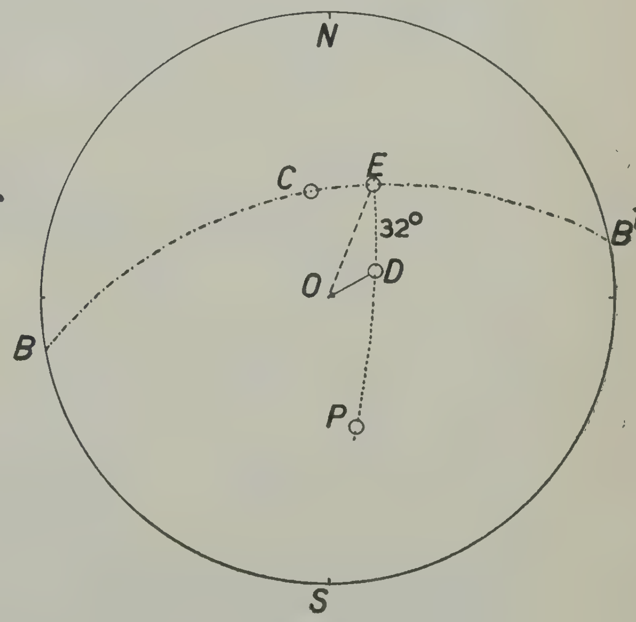


Figure 11



Stereogram 11(c)

(*c*) There is yet another method by which the solution may be attained:

Plot the D.D.H., DO, the bed BCB^1 and the pole of the bed P, in the usual way (stereogram 11 (c)).

As DO and PO are two lines in space through a common point O, only one plane can be drawn through them. Rotate the stereogram until P and D coincide with the same great circle; draw in the latter and note where it intersects the bed (E).

Any great circle through P is normal to the plane BCB^1. Thus the angle between OD and OE (the line of intersection between the planes BCB^1 and EP) measured on the great circle EP is the desired core-bed angle. By reference the latter is given as 32°.

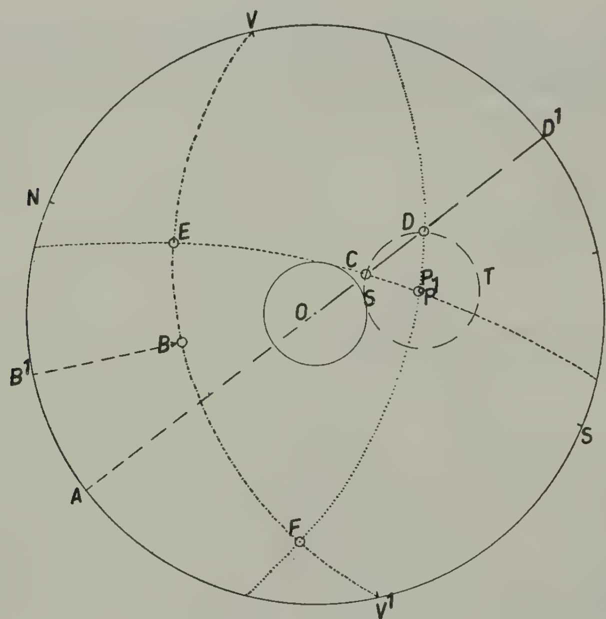
The latter method was also adopted by Walter Bucher in his cyclographic approach to the problem.

Fig. 11 is a perspective diagram illustrating the angle which would appear on the core of an arbitrary D.D.H.

Problem 12

A gold-quartz vein dips 40° in a direction S35°E. It is desired to drill an inclined pilot-hole in a direction N60°W in order to intersect the vein in such a way that a core-vein angle of 70° be recorded.

Question: What is the desired inclination of the hole?



Stereogram 12

Procedure: (*a*) As noted in the previous problem, the core-vein angle is recorded within a plane-normal to the vein. When the vein is horizontal, the angle of inclination of the D.D.H. will be the same as the recorded core-vein angle.

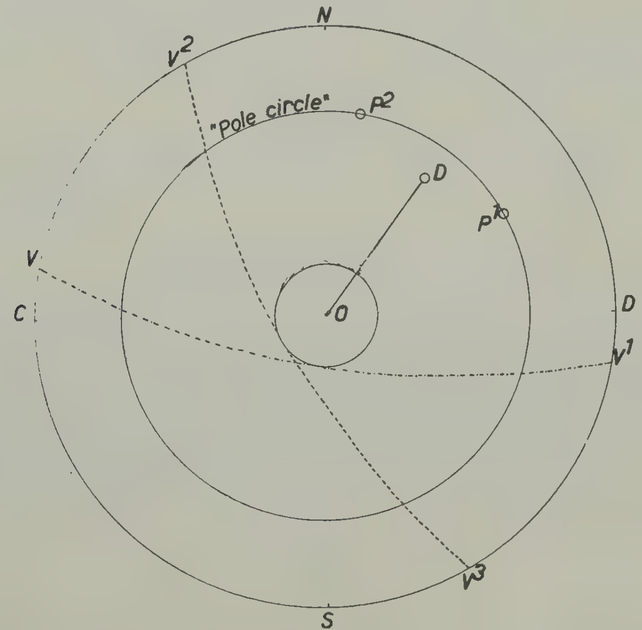
Record the geographical north and south positions on the stereo gram and plot the line VV^1 representing the strike of the vein (normal to S35°E). The great circle VB^1V^1 records the upper half of the plane of the vein when rotated to the horizontal. The D.D.H. will now plot somewhere on the circumference of a circle having a radius of 20° i.e. (90° — 70°) and a centre coinciding with O, the centre of the net. On the other hand it is known that the vein dips 40° in a direction S35°E. To show this, rotate around the axis VV^1 (B^1 being on the left side of the equator) through an angle of 40°. B^1 is transferred to position B, the great circle VBV^1 recording the attitude of the vein. During this rotation all points on the D.D.H. circle are transferred to new positions located 40° to the right and measured along small circles. This must be accomplished by transferring as many points as possible to facilitate construction of the new circle. The D.D.H. point must now appear somewhere on this circle.

Plot the pole P of the vein which is located 90° distant from the point B. Draw the line AD^1 through the centre of the net, representing the direction of plunge of the D.D.H. i.e. N60°W. The trace of the latter direction intersects the transferred circle in C and D. Either CO or DO is the required dip vector of the D.D.H.

To check which possibility meets the requirements, draw great circles through P and C, and P and D; these intersect the plane of the vein in E and F respectively. It will be noticed that the core-vein angle DF is 110°, whereas CE is 70°.

CO is thus the required D.D.H. and displays a plunge of 65°. It must be remembered, however, that here we presuppose orientated core. In the more common case both CO (plunge 65°) and DO (plunge 40°) will give the same result.

(*b*) The problem could be solved very much more quickly if we realise that the D.D.H. must be located on a great circle through P, the distance between the two points being 90° — 70° = 20°. This desired position falls on a circle, but P is not the exact centre of this circle because there is an angular distance increase between every alternate 2 degrees as we move from the centre of the net. The exact centre may be determined by dividing the distance ST. With the centre established, the circle may be drawn in (radius SP^1) and the points of intersection with the trace of the direction of plunge of the D.D.H. (C and D) located. Henceforth the problem is the same as before.



Stereogram 13

Problem 13

A diamond-drill hole record-sheet gave the following information:

D.D.H. inclination 30°. Direction of plunge S36°W.

Vein intersected shows a core-vein angle of 65°.

Actual dip of the vein 70° in some direction north.

Determine: The strike and direction of dip of the vein.

Procedure: The dip vectors of a multitude of veins of unknown strike but similar angle of dip (70°) are represented on the stereogram as a circle with radius of 20°, the centre of the circle coinciding with O, the centre of the net.

The poles of these veins also lie on a circle with a radius of $90° - 20° = 70°$ and centre coinciding with O.

Locate these two circles on the stereogram, measuring the angular radius from the centre and noting "vein-circle" and "pole-circle" on the circumference of the respective circles.

Locate the geographical N-S positions and plot the D.D.H. as DO plunging 30° in a direction S36°W.

The core-vein angle is measured on a great circle through D and a pole on the "pole-circle", but the angle between the latter two positions is given as 90° minus core-vein angle. Rotate the stereogram until this angle $90° - 65° = 25°$ appears between D and a point on the "pole-circle", being careful to record the angle on a great circle of the net. It will be noticed that there are two positions P^1 and P^2 which satisfy these requirements. By placing P^1 and P^2 in consecutive order on the equator the two possible great-circles representing the planes of the veins may be drawn in tangentially to the "vein-circle". These locate the two possible strikes VV^1 and V^2V^3.

The correct answer to the problem is a matter of trial and error in the field, but using the recorded northward dip as a clue, it may safely be surmised that VV^1 is the correct strike (N80°W) and that the correct dip is in a direction N10°E.

Problem 14

In an area characterised by very deep weathering, the following data was obtained from a diamond-drill hole:

D.D.H. inclination 30°. Direction of plunge S36°W.

Core-vein angle 65°.

General strike of veins in this area as determined by other means given as N78°W.

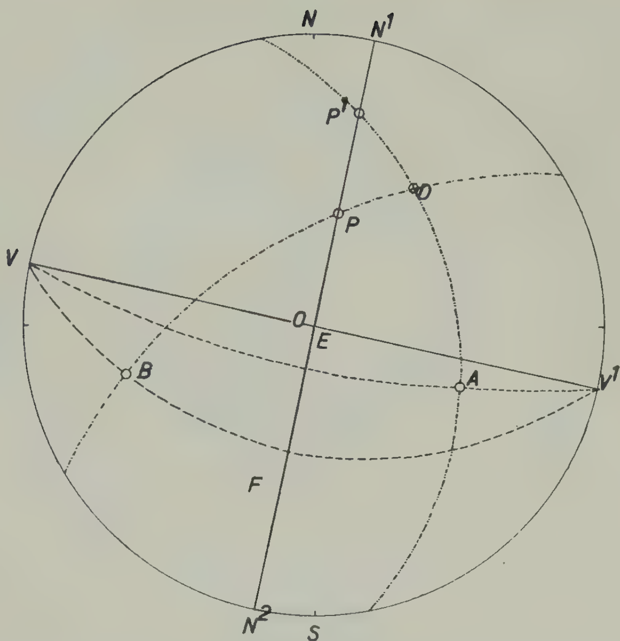
Stereogram 14 on following page.

Question: What is the probable angle of dip of the vein?

Procedure: Plot the geographical positions on the tracing paper. Locate the strike of the vein and indicate it as the line VV^1 through the centre of the net. Draw a line normal to the latter (N^1N^2), simply by utilising the equator and the N-S line.

It is obvious that no matter what the amount of dip, the pole of the vein must fall on the line N^1N^2, which is also a great-circle.

Plot DO, the diamond-drill hole, according to the information given. Rotate the stereogram until a great circle through D intersects N^1N^2 in such a way that the angle between the point of intersection and D equals the complement of the given core-vein angle, i.e. $90° - 65° = 25°$.



Stereogram 14

It will be noticed that there are two great circles conforming to this requirement which cut N^1N^2 in P^1 and P respectively. Measure off 90° on both the great circles PD and P^1D, commencing at the poles in each case and counting south. In this way points B and A can be located. Extend the great circles P^1D and DP so as to include A and B respectively. The two planes VAV^1 and VBV^1 represent two possible positions in space for the vein. With VV^1 coinciding with NS on the net, the respective dips are given as 73° and 42½° in a direction N12°E.

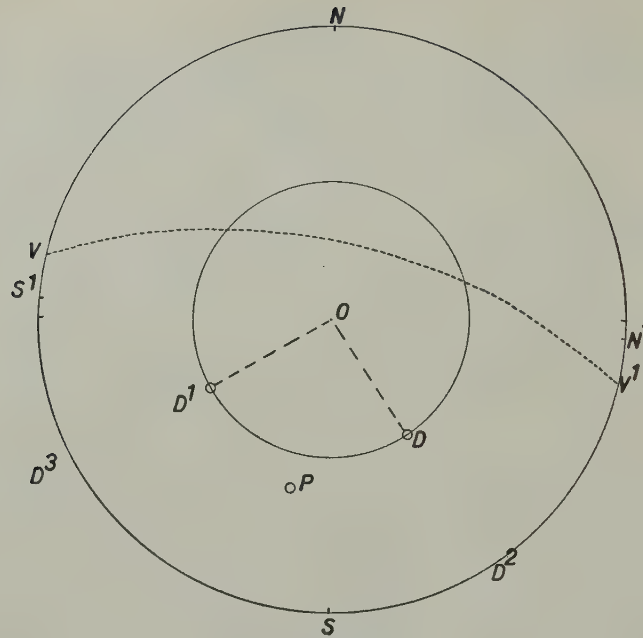
Problem 15

A D.D.H. inclined at 40° in a direction N34°W cuts a vein, and the core displays a core-vein angle of 50°. A vertical D.D.H. cuts the same vein, but in this case the core-vein angle is 29°.
Question: What is the dip and strike of the vein?

Stereogram 15 on following page

Procedure: The information from the vertical D.D.H. is interpreted as proof that the vein has a true dip of 61° (the complement of the core-vein angle). This deduction is based on the fact that there is only one plane through a vertical D.D.H. which is normal to the plane of the vein, and the line of intersection between the two planes defines the direction of dip.

Without fixing any geographic positions on the tracing paper, a plane is plotted having a fixed dip of 61°; direction of strike, however, is arbitrary. This plane re-



Stereogram 15

presents the vein in the problem. Such a plane is presented on the stereogram as VV^1. Construct a circle having a radius of 50° (90° — D.D.H. inclination) and centre O the centre of the net. The inclined D.D.H. must lie somewhere on this circle. Plot the pole P of the arbitrary plane VV^1. Rotate the stereogram until D, a point on the circumference of the circle, and P lie on the same great circle so that the angle $PD = 90 - 50 = 40°$ (the angle between the pole and the D.D.H. equals the complement of the core-vein angle).

It will be noticed that there are two positions D and D^1 conforming to this requirement. Thus either DO or D^1O represents the plot of the D.D.H. which plunges in a direction N34°W. Place DO and D^1O consecutively on the equator and extend them to intersect the circumference of the net in D^2 and D^3 respectively. Now it follows that either D^2O or D^3O is the direction N34°W. Insert the respective north positions for the two cases, i.e. N^1 and N. It is obvious that the strike of the vein VV^1 is either N9°E or N77½°W.

Problem 16

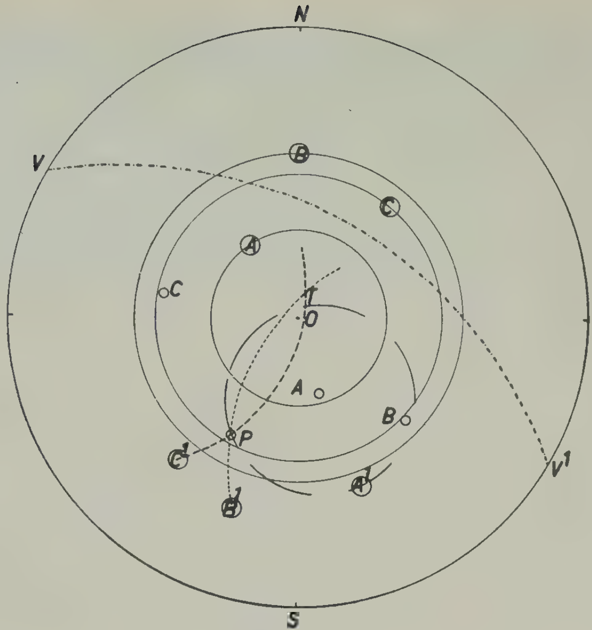
The following information was obtained from three non-parallel diamond-drill holes, which explored a hidden bed.

	Direction.	*Inclination.*	*Core-bed angle.*
No. 1 Hole (AO)	N16°W	60°	57°
No. 2 Hole (BO)	N47°W	36°	31°
No. 3 Hole (CO)	S80°E	40°	38°

Questions:

(*a*) What is the attitude of the bed?

(*b*) Are these the only D.D.H's which would display core-bed angles as given?



Stereogram 16(a)

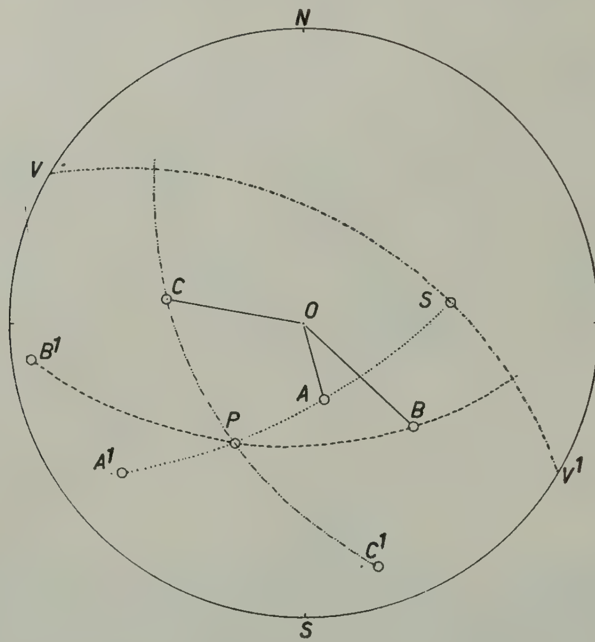
Procedure: (*a*) The position of the pole of an unknown bed must fall somewhere on the circumference of a circle with radius 90° minus core-bed angle and with centre the plotted position of the D.D.H.

On the tracing paper plot the D.D.H. No. 1 as AO; also plot the geographical N-S positions accordingly. With OA on the equator rotate through 30° around an axis normal to it, to transfer the D.D.H. to a vertical position. With O as centre draw a circle with radius $90° - 57° = 33°$. Restore OA to its original position by rotating back through 30°. During the latter step the pole-circle of AO will transfer to a new position, conforming to an arc through A^1. Repeat the same procedure for BO and CO, so as to obtain pole-arcs through B^1 and C^1 respectively. All these pole-arcs intersect in a common point P. The latter is the required pole of the bed. With P on the equator, locate a point 90° away along the equator and draw a great circle through it. The latter is the required trace of the bed (VV^1) which dips 50° in a direction S30°W.

Referring to stereogram 16 (*a*), it will be noticed that the whole solution is presented as one. This is done only for the sake of clarity. It must be realised that the construction of the individual pole-arcs involves three different amounts and directions of rotation, which must affect the stereogram as a whole if considered as a unit. The three constructions may be done on separate stereograms or, if presented as one, all the elements, except the pertinent D.D.H. and its pole circle, must be ignored.

Furthermore the respective transferred pole-circles are only shown in part. This is to eliminate confusion which may arise from the presence of too many lines. Practice will show the required side of the pole-circle to be transferred. Extreme care should be taken when transferring these pole-circles, as a small triangle of error may develop at the assumed point of intersection, P. The importance of this factor becomes

clear when we realise that the transferred pole-circles very nearly intersect in another point T. This point actually simulates the required pole. This may occasionally happen when core-bed angles are nearly similar, as in the case of D.D.H. No. 3 and No. 2.



Stereogram 16(b)

(*b*) In stereogram 16 (*b*) the bed VV^1 and the three D.D.H's AO, BO and CO are reproduced. Furthermore, the great circles PA, PB and PC, through the common pole P, are also shown. Referring to AO, it becomes clear that the angle AS (57°) is the required core-bed angle. But if the great circle PS be extended to its limit on both sides, and an angle of 57° is measured off from S in the opposite direction to SA, the position A^1 is located. The measurement of this angle is accomplished in two steps. If measurement were towards the north pole of the net, the latter would be reached after a count of $42\frac{1}{2}$°. After an imaginary count of 180° this position will reappear at the south pole; count another $14\frac{1}{2}$° on the same great circle to locate point A^1. Similarly locate positions B^1 and C^1. Now it becomes clear that the angles PA^1, PB^1 and PC^1 equal the complements of the respective core-bed angles. Thus A^1O, B^1O and C^1O represent three other D.D.H's having respective plunges of 13°, $4\frac{1}{2}$° and 9° in directions N51°E, N83°E and N16°W. These also conform to the requirements for core-vein angles as specified.

Problem 17

A contact-drift located on the footwall of a mineralised dyke which trends N80°E dipping 55°, S10°E, intersects a fault which strikes N10°W and dips 70°,

N80°E. A cross-cut at the fault intersection displayed slickensides on the face of the fault trending N25°E (Fig. 17).

Questions:

(*a*) Where is the displaced portion of the dyke located relative to the drift?

(*b*) What is the required inclination of a D.D.H. in a direction N25°E, so as to intersect the displaced portion.

(*c*) If the displaced block be intersected at 100′ in a D.D.H. plunging 57° in a direction N25°E, how far must the cross-cut be extended in a direction N10°W in order to intersect the dyke.

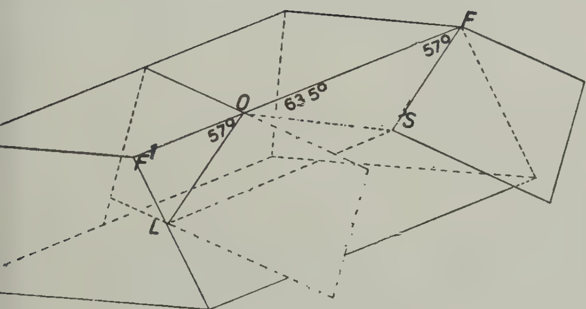
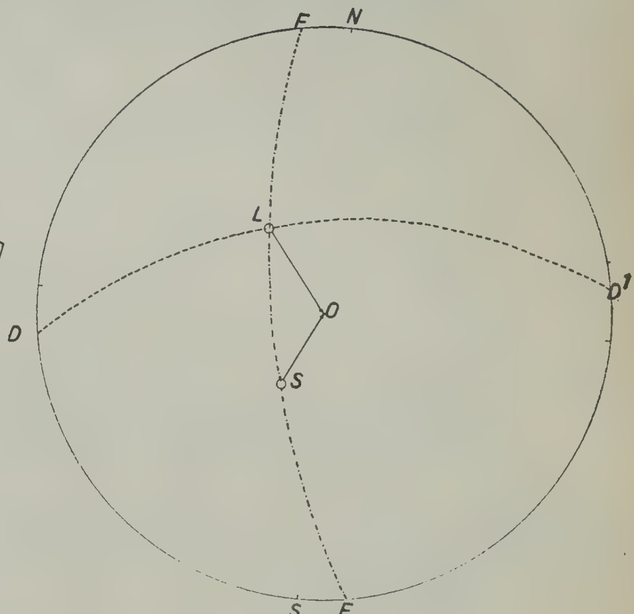


Figure 17



Stereogram 17

Procedure: Plot the fault plane (FF^1) and the footwall-contact plane of the dyke as DD^1. Also plot the direction of the slickensides; the latter intersects the fault plane in S. From the stereogram the following information may be readily deduced:

(1) The line of intersection between the fault and the dyke is LO; it plunges 52° in a direction S37°E.

(2) The angle of pitch of LO on the fault plane is 57°.

(3) The lineation (slickensides) on the fault pitches $63\frac{1}{2}°$ and its plunge is 57°.

(*a*) and (*b*). The displaced portion of the dyke is located to the north of the drift and will be located by a D.D.H. plunging 57° in a direction N25°E.

(*c*) In Fig. 17 the angle FOS equals the pitch of the slickensides, and the angle OFS is the angle of pitch of the line of intersection between the fault and the dyke.

$$\left.\begin{array}{l}\angle \text{FOS} = 63\tfrac{1}{2}° \\ \angle \text{OFS} = 57°\end{array}\right\} \text{Thus angle OSF} = 180° - 120\tfrac{1}{2}° = 59\tfrac{1}{2}°$$

From trigonometry it follows that:

$$\frac{\sin\ 59\frac{1}{2}^\circ}{\text{OF}} = \frac{\sin\ 57^\circ}{\text{OS}} = \frac{\sin\ 57^\circ}{100}$$

$$\therefore\quad \text{OF} = \frac{100 \times \sin 59\frac{1}{2}}{\sin 57^\circ} = 102{\cdot}8'$$

The reader is reminded that the distance OS = 100′ was recorded by the diamond-drill hole, and is the net slip along the fault; OF being the strike slip. Thus the cross-cut will have to be extended for another 102·8′ to the north.

Throughout this problem it was assumed that the slickensides recorded total movement along the fault.

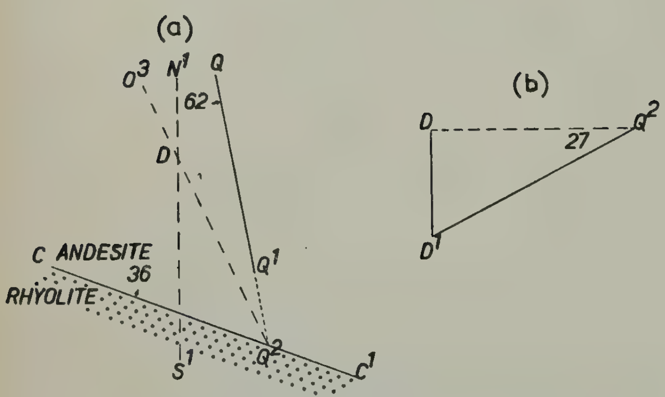
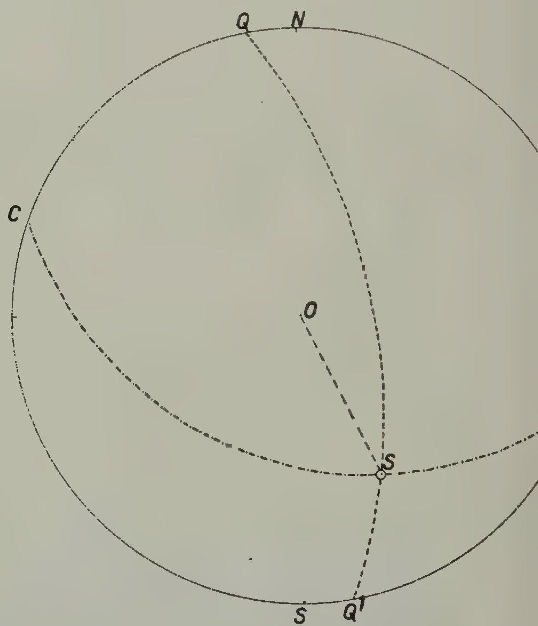


Figure 18



Stereogram 18

Problem 18

A certain mining area is characterised by chalcocite lodes in the vicinity of the intersection between quartz-veins and rhyolite-lava sheets, the latter always overlain by andesite. The above diagram is a map of part of this area, the various elements being indicated (Fig. 18 (*a*)).

Question: It is required to explore the possibilities of ore at depth by a vertical D.D.H. at a suitable locality somewhere along the grid line N^1S^1. At what depth will this hole intersect ore?

Procedure: Transfer the strike of the quartz-vein QQ^1 (N10°W) and that of the andesite-rhyolite contact CC^1 (S70°E) to the tracing paper. In this way the north and south positions of the stereogram will be established. Regard each as a consistent plane in space, and procure the cyclographic projection from the dip evidence as given. The two planes intersect along line SO, which plunges 27° in the direction N$25\frac{1}{2}$°W.

On the map (Fig. 18 (*a*)) extend the quartz-vein QQ^1 to intersect the rhyolite-andesite contact in Q^2. Transfer the direction of plunge of SO from the stereogram to the map, using Q^2 as reference point. Thus Q^2Q^3 represents the horizontal projection of the line of intersection, and also a possible linear ore-body along the vein-rhyolite intersection.

The required position for the D.D.H. will be at D, where the north-south grid line N^1S^1 is truncated by the projection of SO.

The map was ideally chosen to represent a horizontal surface, thus D and Q^2 have the same elevation. A simple mathematical relation between the depth of the D.D.H. and distance DQ^2 is derived from Fig. 18 (*b*).

$$\text{i.e.}\quad \tan 27°\ (\text{plunge of SO}) = \frac{DD^1}{DQ^2}$$

The distance DQ^2 can be scaled.

$$\therefore\ DD^1\ (\text{depth}) = DQ^2 \times \tan 27°.$$

Problem 19

Let us assume that the D.D.H. in problem 18 intersected ore, and that instead of the case being clear-cut, matters are complicated by a fault which has displaced both the vein and the rhyolite contact, as shown in the accompanying map. Use the same information as in problem 18 (Fig. 19).

Questions:

(*a*) In what direction must drilling proceed on the south side of the fault?

(*b*) What is the net slip, plunge and pitch on the fault?

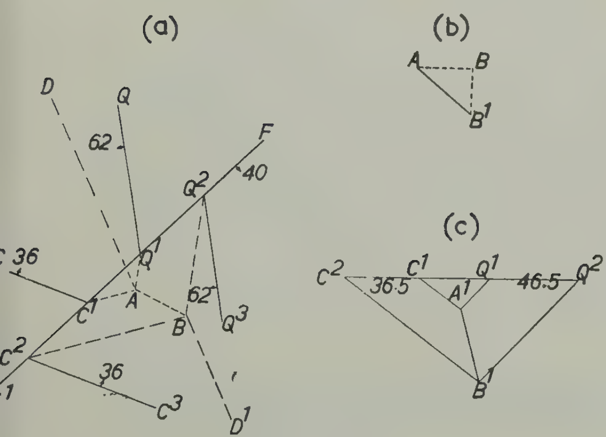
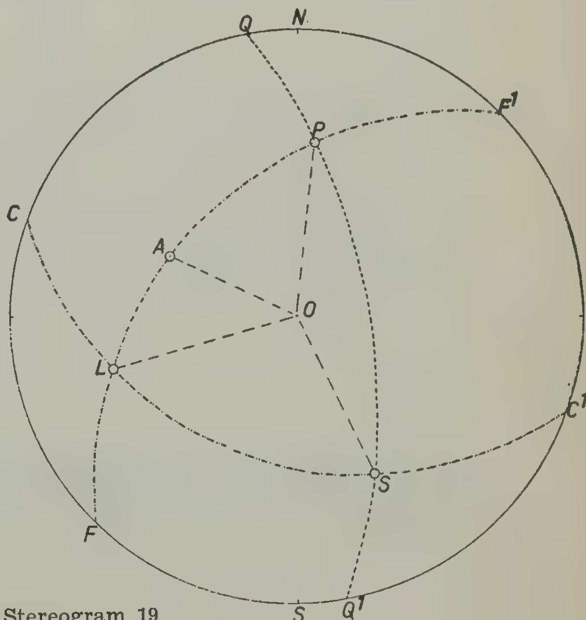


Figure 19



Stereogram 19

Procedure: Obtain the cyclographic projection of the fault plane FF^1 on stereogram 18. The strike is given as N45°E; dip 40°SE. From the stereogram the following information is readily ascertained:

(1) LO is the trace of the rhyolite-andesite contact on the fault plane; it plunges $22\frac{1}{2}°$ in a direction N74°E, its pitch within the fault plane being $36\frac{1}{2}°$.

(2) PO, the trace of the quartz-vein on the fault-plane, plunges 28° in a direction S6°W; its pitch being $46\frac{1}{2}°$ within the plane of the fault.

A "bird's-eye" view of the fault makes it obvious that the trace of the rhyolite-andesite contact and the quartz-vein on the fault, is twofold; it appears on the footwall as well as on the hanging wall of the latter.

Transfer the direction LO and PO to the map so that L coincides with C^1, and C^2 and P with Q^1 and Q^2. The two traces C^1A (LO) and Q^1A (PO) meet in A; similarly C^2B (LO) and Q^2B (PO) intersect in B. Thus the rhyolite-vein intersection will pierce the footwall of the fault in A and the hanging wall side of the fault in B. Now AB must represent the horizontal projection of the net slip along the fault. For clarity transfer the projection of the ore-body as obtained from solution 18 to both A and B (DA and BD^1).

Measure (or transfer directly) the direction AB and transfer to stereogram 19. This direction BAO cuts the fault plane in A. The line AO plunges $38\frac{1}{2}°$; its pitch on the fault plane being 75°.

Answers:

(*a*) Proceed the drilling programme along line BD^1 (map).

(*b*) The distance AB (map) is the horizontal projection of the net slip. From the stereogram it (AO) plunges $38\frac{1}{2}°$.

$$\text{Thus:} \quad \cos 38\tfrac{1}{2}° = \frac{AB}{AB^1} \quad \text{(measure AB directly from map)}$$

$$\therefore \quad \text{net slip } AB^1 = \frac{AB}{\cos 38\frac{1}{2}°} \qquad \text{(Fig. 19 }(b))$$

Alternative method for solution of (*b*):—

$(b)^1$ From Fig. 19 (*a*) Q^2C^2B represents the projection of the fault plane. The angle of pitch of C^1A equals that of $C^2B = 36\frac{1}{2}°$, and that of Q^1A equals that of $Q^2B = 46\frac{1}{2}°$.

Measure off the distance between C^2 and C^1, C^1 and Q^1, and Q^1 and Q^2 from the map and construct graphically the appearance of their linear elements on the plane of the fault (Fig. 19 (*c*)).

From this it is seen that the distance A^1B^1 is the desired net slip.

Problem 20

The following map shows a rotational fault displacing a mineralised dyke. This adjacent area is characterised by E-W trending dykes, which suggest that the east side participated in the faulting (Fig. 20).

Question: Through what angle was the rotation along the fault accomplished?

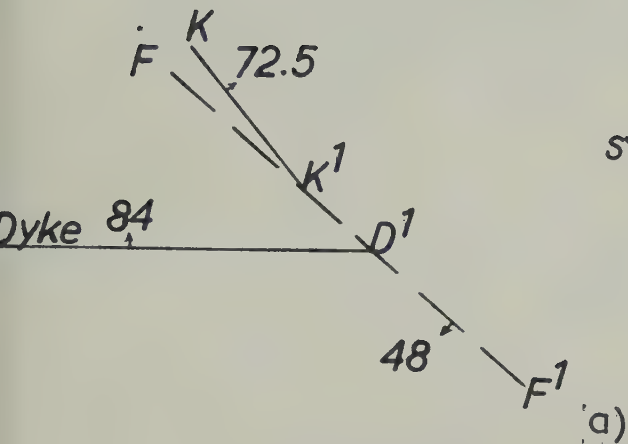
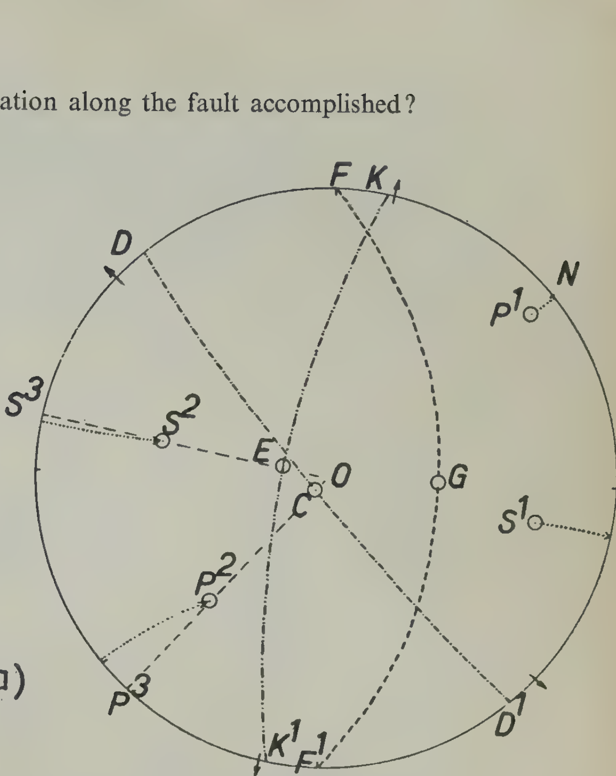


Figure 20



Stereogram 20(a)

Procedure: On tracing paper plot the attitude of the two portions of the dyke (DCD^1 and KEK^1). Also plot the fault as FGF^1.

Say the fault was a horizontal one; then rotation within the plane of the fault (around an axis normal to it) could have accomplished only a variation in the strike of the dyke, but the amount of dip would have remained the same.

With FF^1 in the N-S position of the net, rotate through 48° to the right (around FF^1 as axis) in order to transfer the fault to a horizontal position. During the process P^1 and S^1, the respective poles of DCD^1 and KEK^1, transfer to the new positions P^2 and S^2 (as indicated by the dotted lines and arrows).

Locate the strike lines of the transferred portions of the dyke, of which P^2 and S^2 are the poles. The positions of intersections of these lines at the circumference of the net are indicated by four arrows. Measure the acute angle between the arrows on the circumference of the net ($59\frac{1}{2}$°).

Answer: The rotation along the fault was accomplished through an angle of $59\frac{1}{2}$°.

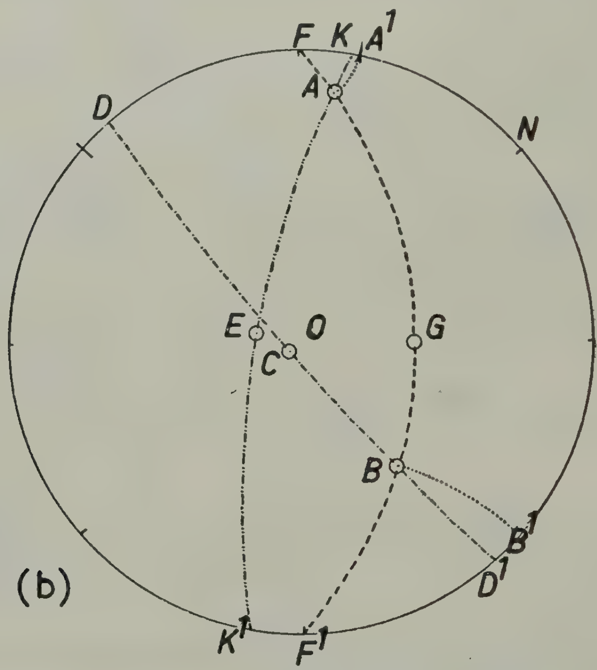
Alternative method:

(1) With the pole P^2 on the equator, extend OP^2 to the circumference of the net, arriving at position P^3. Similarly locate S^3 by extending OS^2 to the circumference of the net.

The angle S^3P^3 is the required angle of rotation.

(2) During rotational faulting the line of intersection between the dyke and the fault plane suffers displacement, but if the fault is rotated to a horizontal position,

the two lines of intersection (the original one and the displaced one) must attain a horizontal position because they occur on the fault plane. Thus these lines of intersection will become the lines of strike of the two portions of the dyke, when the fault is rotated to a horizontal position.



Stereogram 20(b)

In stereogram 20 (*b*) the fault FGF^1 and the two portions of the dyke DCD^1 and KEK^1 are again reproduced. Note the line of intersection between the fault and DCD^1 (BO) and KEK^1 (AO).

Rotate through 48° around axis FF^1 so as to bring the fault to a horizontal position. Position A and B move along small circles to reach positions A^1 and B^1. The obtuse angle between the latter points is $120\frac{1}{2}°$; thus the acute angle must be $59\frac{1}{2}°$.

If stereograms 20 (*a*) and 20 (*b*) are superimposed, it becomes evident that OA^1 and OB^1 coincide with the directions of strike as obtained on stereogram 20 (*a*).

Bucher solved a similar type of problem by using the plane-method only. Jerome Fisher criticised the latter, and showed that the problem could be solved by using the P.P. method. But he pointed out that the P.P. method is only applicable if the angle between the undisturbed portion and the faulted portion of the bed or dyke equals the angle of rotation.

For the simple reason that the P.P. method suffices only in exceptional cases, the S.P. method, as outlined on stereogram 20 (*a*) and under alternative method (2), is deemed more practical in solving problems of this type.

Problem 21

An inclined shaft plunging 40° east is sunk for the purpose of developing an ore-body which pitches 36° in an easterly direction within the confines of a norite dyke which dips 64°, N30°E. The direction between the surface exposure of the ore-body and the shaft collar is given as N10°E; the surface being horizontal (Fig. 21 (*a*)). *Question:* It is desired to check the tenor of the ore-body underground from a drill-site in the shaft. How can this drilling be accomplished in the shortest possible time with one D.D.H.? Where is the hole located?

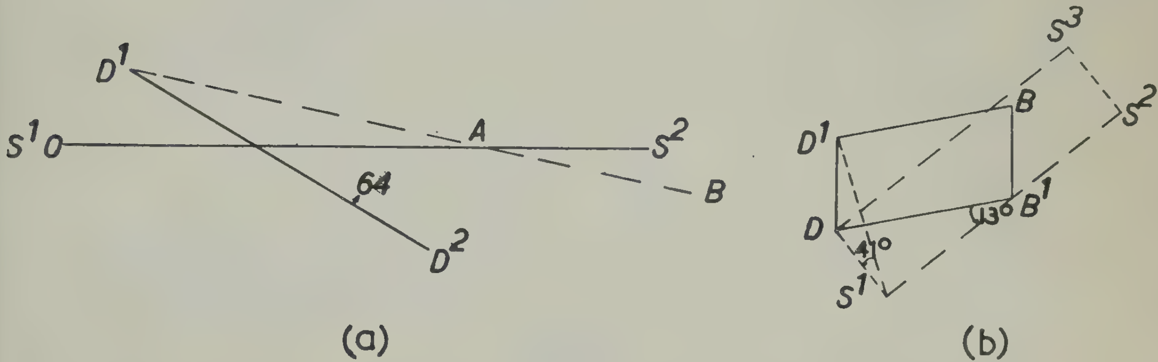


Figure 21

Procedure: The following plan (Fig. 21 (*a*)) shows the horizontal projection of the shaft S^1S^2, S^1 being the entrance; and the dyke D^1D^2 with position D^1 as the outcrop of the ore-body. (For practical purposes, the ore-body, which is linear in space, may be regarded as a point on exposure).

To answer the first part of the question, it is necessary to realise that the desired D.D.H., requiring the least expenditure and time, must be the same as the shortest possible hole that can be drilled from the shaft to the ore-body. If the ore-body and the shaft are visualised as lines in space, it becomes obvious that the shortest possible line between them is one standing normal to both.

Stereograms 21 (a) and 21 (b) on following page

Now proceed as follows: (Stereogram 21 (a)):

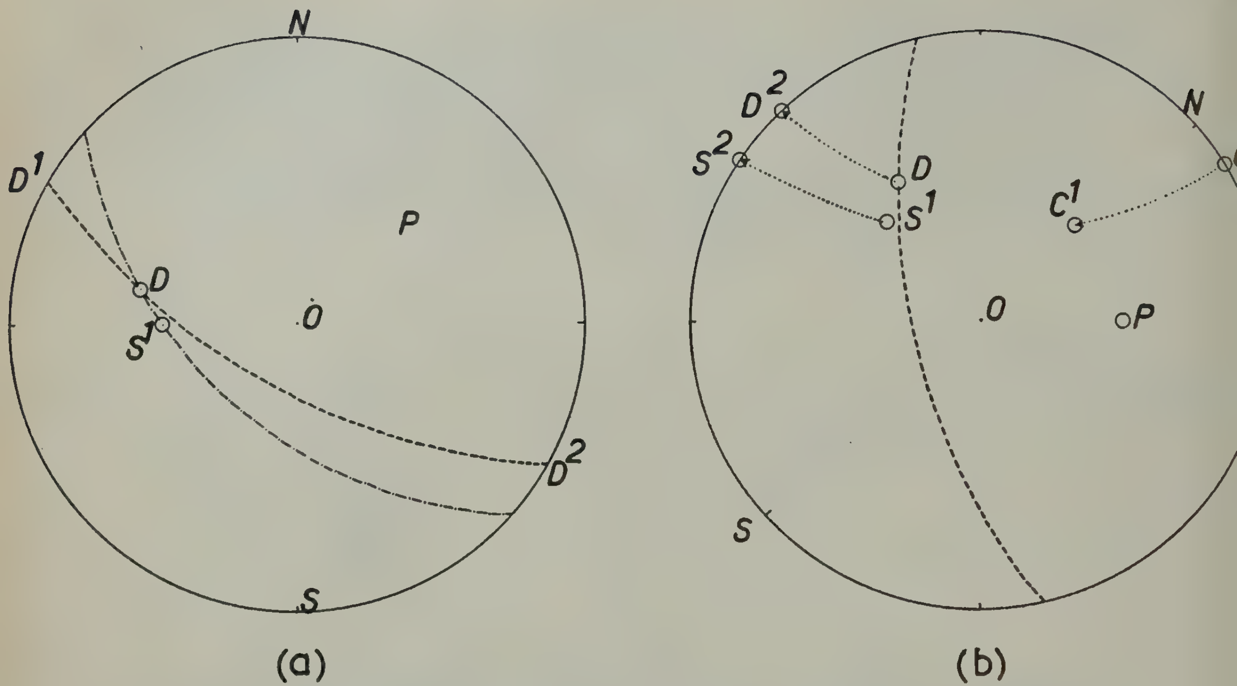
(*a*) On tracing paper plot the dyke as D^1D^2. With D^1D^2 in the N-S position locate position D, so that the angle D^1D equals 36° (angle of pitch of ore-body).

(*b*) DO represents the ore-body, and by reference it plunges 32° in a direction S77½°E.

(*c*) Plot the shaft on the stereogram as line S^1O plunging 40° east.

(*d*) Draw the great circle through D and S^1 and locate its pole P, 90° away.

PO represents a line in space which is normal to the shaft as well as to the linear ore-body. From the stereogram the plunge of PO is given as 39° in a direction S43°W. But, because the ore-body is clearly always above the shaft, the D.D.H. must operate in the opposite direction. Thus the required D.D.H. is inclined + 39° (up) in a direction N43°E.



Stereogram 21

To answer the second part of the question, we require the following explanation:

Two non-parallel lines lying respectively in two horizontal planes (one above the other) will show a point of intersection on a horizontal plane of projection. The shortest distance between the two lines is a vertical line whose projection (on the same horizontal plane of projection) coincides with the point of intersection referred to. Thus point A (Fig. 21 (*a*)) the point of intersection between the trace of the ore-body and shaft is not the projection of the required D.D.H., because it is known that the D.D.H. is inclined. The solution of the problem is accomplished by transferring the D.D.H. to a vertical position and thus rotating the ore-body and shaft to horizontal positions.

Procedure: With P on the equator rotate through 51° to transfer the D.D.H., PO, to a vertical position (stereogram 21 (*b*)). During this rotation the shaft S^1O and the ore-body DO attain the respective horizontal positions S^2O and D^2O. The new trend of these two linear elements is given as S77°W and E-W.

How does this operation affect Fig. 21? It is obvious that S^1 and D^1 are fixed positions on the shaft and ore-body respectively and thus rotation will disturb their positions relative to one another.

Plot the direction S^1D^1 on the stereogram 21 (*a*) as the line OC. When PO is transferred to a vertical position, position C moves along a small circle to C^1; C^1O now plunges 41° in a direction S2°E. Thus D^1S^1 is not horizontal any more (Fig. 21(*a*))

but plunges in the direction indicated above. Fig. 21 (*b*) represents a view of the position in space of the ore-body and shaft with the D.D.H. in a vertical position.

D^1B (ore-body) lies parallel to the plane $DS^1S^2S^3$.

$DS^1S^2S^3$ contains the shaft S^1S^2.

DD^1 and BB^1 are normal to both BD^1 and DB^1.

D^1S^1 is the new position in space of the line connecting the exposure of the ore-body (D^1 in Fig. 21 (*a*)) and the entrance to the shaft.

From triangle DD^1S^1

$$\cos 41^\circ \text{ (angle of plunge of } D^1S^1) = \frac{DS^1}{D^1S^1}$$

From triangle S^1DB^1

$$\tan 13^\circ \text{ (angle between } S^2O \text{ and } D^2O \text{ on stereogram 21 } (a)) = \frac{DS^1}{S^1B^1}$$

$$\therefore \quad S^1B^1 = D^1S^1 \times \frac{\cos\ 41}{\tan\ 13} = X$$

(D^1S^1 measured on map Fig. 21 (*a*)).

The distance X is the required distance from the entrance of the shaft to where the D.D.H. should be drilled.

The length of the D.D.H. is given as $D^1S^1 \times \sin 41^\circ$.

Problem 22

In a certain mining area mineralised dykes are characterised by unusually uniform hanging wall contacts.

The following map (Fig. 22 (*a*)) shows such a dyke.

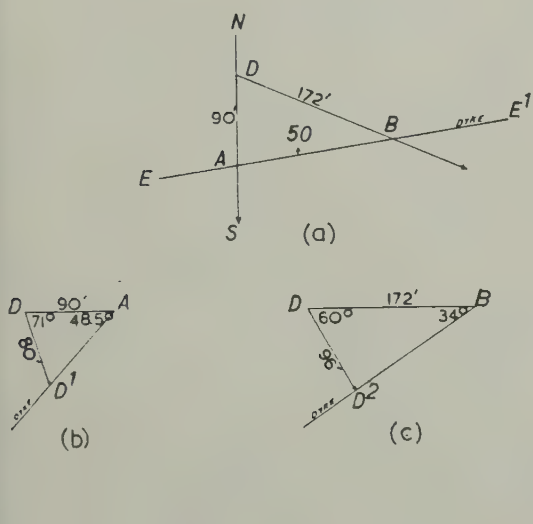
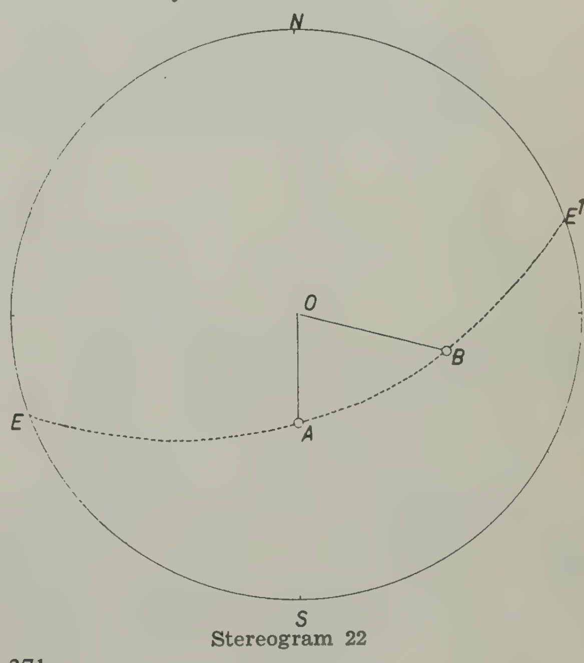


Figure 22



Stereogram 22

It is desired to check the downdip extention of the dyke with two inclined holes, one plunging south and the other plunging S66°E from position D.

Questions:

(1) If the southerly plunging hole intersected the dyke at a depth of 80′, what was its plunge?

(2) The other hole was inclined at 60° (S66°E). Where did it intersect the dyke?

Procedure: On tracing paper plot the dyke as EE^1. Locate the apparent dip of the dyke in the two directions north and N66°W, i.e. opposite to the directions in which the D.D.H's plunge. This is given on the stereogram as AO ($48\frac{1}{2}°$) and BO (34°).

Answers (1) *and* (2): Construct graphically two triangles using the apparent dip angles as recorded on the stereogram and the known distance DA and DB from the plans (Figs. 22 (*b*) and 22 (*c*)).

The D.D.H. which plunges in a southerly direction has an angle of plunge of 71°. The D.D.H. plunging 60°, S66°E intersected the hanging wall of the dyke at 96′.

In the above problem the surface was taken as horizontal. The problem may be solved by allowing for elevation differences between the D.D.H. site and the dyke in the graphical solution, if the surface is not horizontal.

Problem 23

The following map indicates a mineralised dyke truncated by a fault; the surface of the outcrops being essentially horizontal (Fig. 23 (*a*)).

Question: What would be the attitude of the dyke and the fault on a vertical north-south section indicated as the line N-S on the map?

Procedure: Plot the dyke as DD^1 and the fault as FF^1 on the tracing paper (Stereogram 23). Draw in the line connecting the geographic north-south poles of the stereogram Locate the points of intersection between this line and the fault (A) and the dyke (B).

AN (59°) and BS (50°) are the respective plunges of the fault and of the dyke on the vertical N-S plane (Fig. 23 (*b*)).

Question: If a D.D.H. be drilled from position N (see plan) inclined 44° in a direction S20°W, where would it intersect the two portions of the dyke on both sides of the fault. Also indicate where it would intersect the fault.

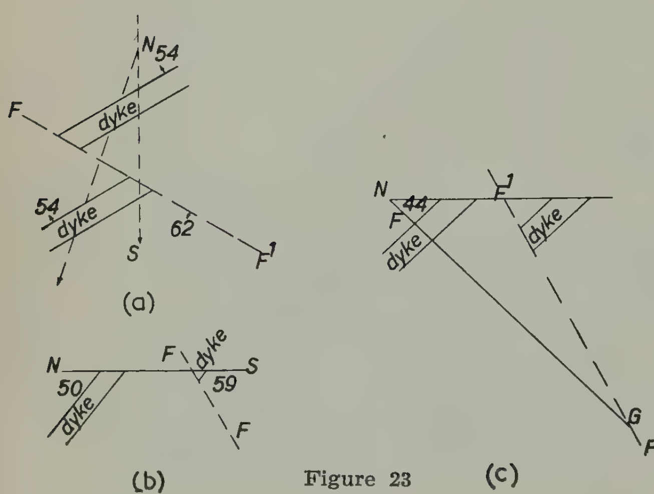
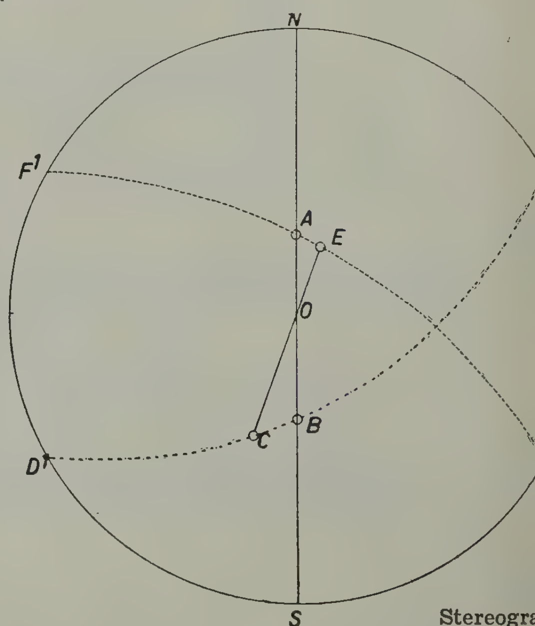


Figure 23



Stereogra

Procedure: Procure the apparent dip of the dyke and fault as seen in a vertical section trending S20°W, using the method outlined above. The apparent dip of the dyke is 90°—CO (42°) and that of the fault 90°—EO ($61\frac{1}{2}$°). (Stereogram 23.)

This information has been used to find the solution graphically as in Fig. 23 (*c*). As indicated, the D.D.H. will cut the footwall-dyke at F and the fault in G (distance in feet can be scaled off). It failed, however, to intersect the hanging wall dyke, as the angle of the hole was miscalculated.

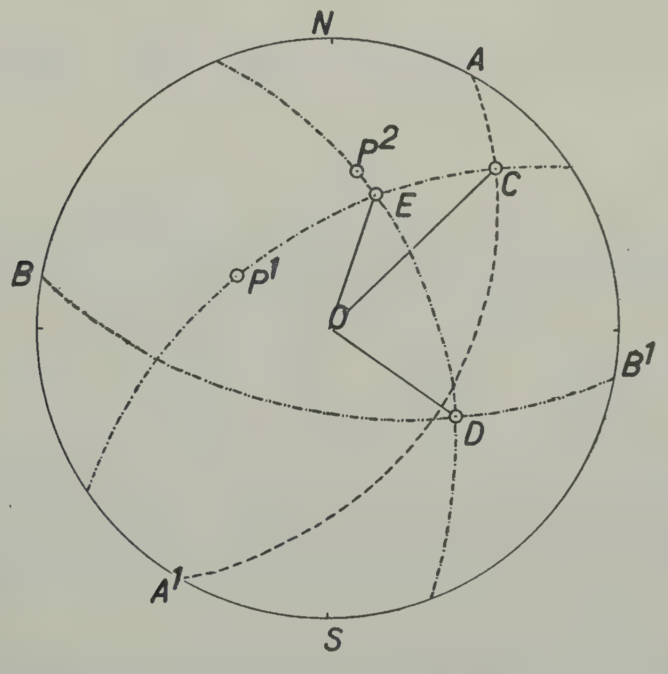
This problem clearly shows how valuable information on fault problems may be obtained by using the S.P. method.

Problem 24

A certain mining area is characterised by cylindrical-shaped replacement ore-bodies in limestone. The rake (trend) of the individual ore-pellets usually parallels that of the bodies proper. The following data was obtained while mapping in an area which exposed good mineralization confined to an anticline:—

The two flanks of the anticline (AA^1 and BB^1), have respective dips of 40° and 58° in the direction N60°W and N10°E respectively. The arrangement of the long axis of the elliptical ore-pellets collectively produce a strong lineation. The average pitch of the lineation on the two flanks of the fold was determined as 20° SW and 40° NW.

Determine: The plunge of the ore-pellets and thus the attitude in space of the cylindrical ore-body.



Stereogram 24

Procedure: The solution of the problem is based on the following concept. The intersection between a cylinder (in space) and a plane which is neither parallel nor normal to the cylinder is an ellipse. The long axis of the ellipse corresponds to the line of intersection between this plane and one normal to it—the latter also containing the axis of the cylinder.

On tracing paper superimposed on the net obtain the cyclographic projection of the two flanks of the fold as planes AA^1 (dip N60°W) and BB^1 (dip N10°E). With position A on the north pole of the net count off 20° south along the trace of plane AA^1 locating position C. The line CO now corresponds to the lineation as recorded on one flank of the fold. Similarly locate DO (the lineation on the other flank of the fold) counting 40° north while holding B^1 on the south pole of the net. Now locate the respective poles P^1 and P^2 of the planes AA^1 and BB^1.

A great circle through P^1 and C intersects plane AA^1 along the lineation CO Thus CO must be the projection on to the plane AA^1 of the long axis of the cylindrical ore-pellet. On the other hand this axis must also project somewhere within the plane P^1C on the stereogram. Similarly the great circle through P^2 and D intersects plane BB^1 along the lineation DO. Because it is assumed that the attitude of the ore-pellets in space remains the same, it follows that the axis must also project somewhere within the plane P^2D.

The two great circles P^1C and P^2D intersect in a point E. Thus EO is a line in space common to both, and it follows that EO corresponds to the attitude of the ore-pellets. By reference EO plunges 38° in a direction S20°W.

The required plunge of the assumed cylindrical ore-body is given as 38°, S20°W.

K. E. Lowe (1956) used more or less the same principles to obtain the lineation in the Storm King granite, but he adopted the Bucher projection instead.

Problem 25

The accompanying plan (Fig. 25 (*a*)) shows a bed truncated by two normal faults. The later fault has displaced the earlier one. Field evidence indicates that movement along both faults was down the dip of the fault plane.
Determine: The vertical displacement (in feet) along both faults.

		Strike.	*Dip.*	
Data:	Bed (CC)	N60°E	30°	N30°W
	Fault (AA)	N50°W	60°	S40°W
	Fault (BB)	N20°E	45°	S70°E

Procedure: On tracing paper obtain the cyclographic projection of the bed and both faults as CC, AA and BB respectively. Locate the line of intersection (DO) between the later fault BB and the bed CC on the stereogram. Similarly locate the line of intersection (EO) of the same fault and the earlier fault AA. By reference find the angle of pitch of the two lines of intersection on the fault plane BB.

That is, DO pitches 20° and EO 55°.

Fig. 25 (*b*) is a section parallel to the fault plane BB, and the various points of emergence of the bed and fault AA are shown to scale. When the fault plane BB is viewed its intersection with the bed and fault AA will appear as lines on the hanging

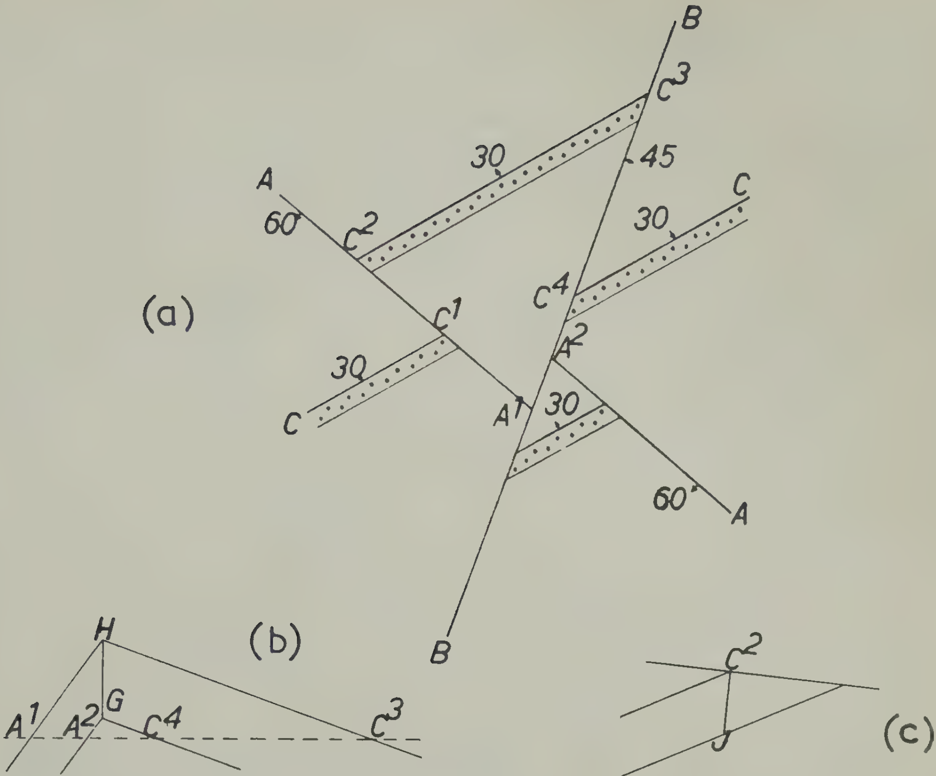


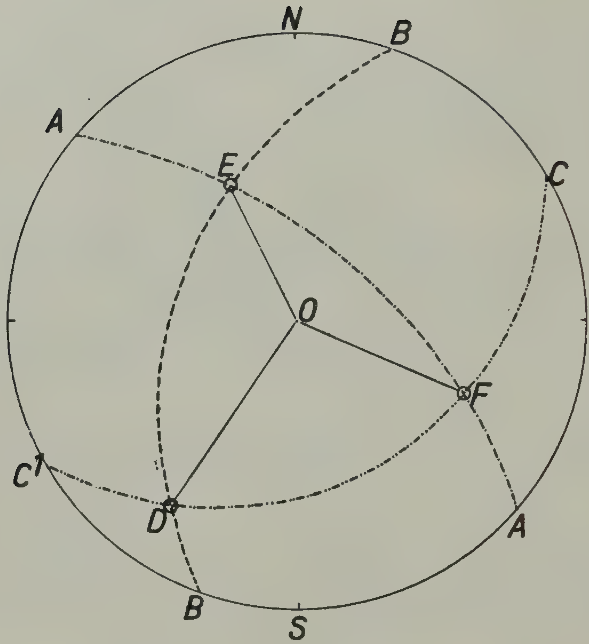
Figure 25

wall as well as on the footwall. These lines are inserted as shown, because the pitch angles DO and EO (stereogram) are known. When extended the hanging wall and footwall intersections meet in G and H. The dipslip along the fault BB is given as HG and the distance may be scaled off (141·4 feet).

A simple trigonometrical function shows that the vertical displacement on fault BB equals $141 \cdot 4 \times \sin 45° = 100$ feet. To answer the second part of the question, proceed as follows: On the stereogram locate the line of intersection between fault AA and the bed (FO). FO pitches 29° on the fault plane AA. Fig. 25 (*c*) is a section parallel to the plane of the fault AA. As before, the line of intersection between the fault plane and the bed will again appear on the hanging wall and footwall. Construct the two lines as shown, using the pitch angle of 29°. A line normal to the horizontal trace of the fault AA pierces the latter in C^2 and the line of intersection on the hanging wall in J. The distance C^2J is the dipslip along the fault AA and measures 115·5 feet.

The vertical displacement along fault AA is given as $115 \cdot 5 \times \sin 60° = 100$ feet.

Problem 25 is not the most difficult case that may manifest itself in areas characterised by normal or reverse faulting. But the method remains the same no matter how many intersecting faults are involved. To compare the method outlined here with the standard methods involving contour lines, the reader is referred to an admirable article by George Dickinson (1954).



Stereogram 25

Problem 26

A vein strikes north-south and dips 60° east. An ore-shoot confined to the vein plunges 50° in a direction N44°E. It is desired to check the grade of the ore from the surface by two inclined D.D.H's, but these holes must be drilled in a plane which contains the ore-shoot DCD^1 as well as being normal to the vein. One hole plunges in a direction N8°E and the other forms an angle of 32° with the former (Fig. 26).

Questions:

(*a*) What is the position in space of the plane in which diamond-drilling must be accomplished?

(*b*) What is the plunge of the first hole?

(*c*) What is the plunge and direction of the last hole?

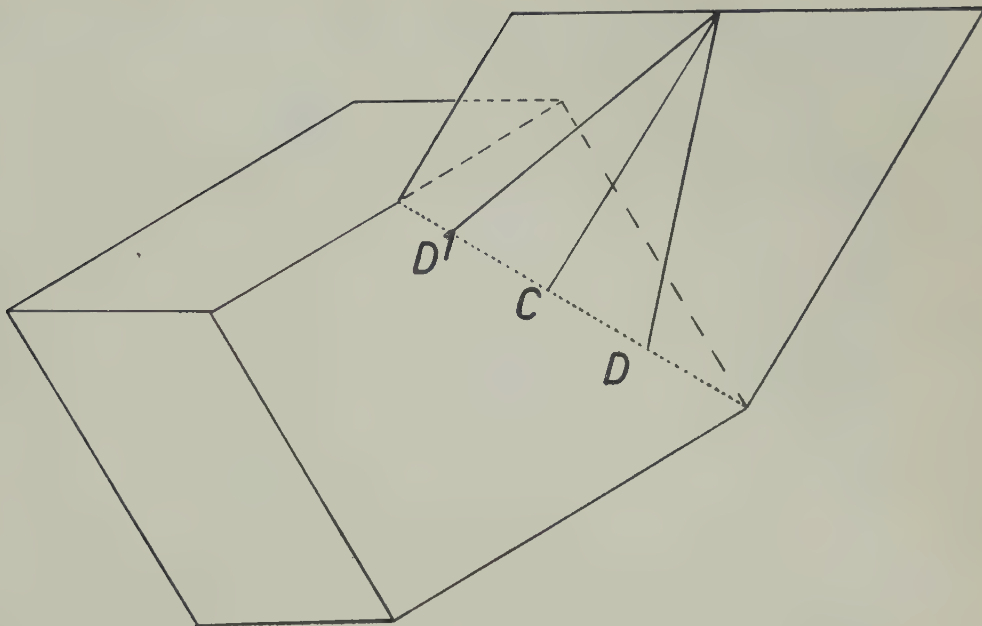


Figure 26

Stereogram 26 on following page.

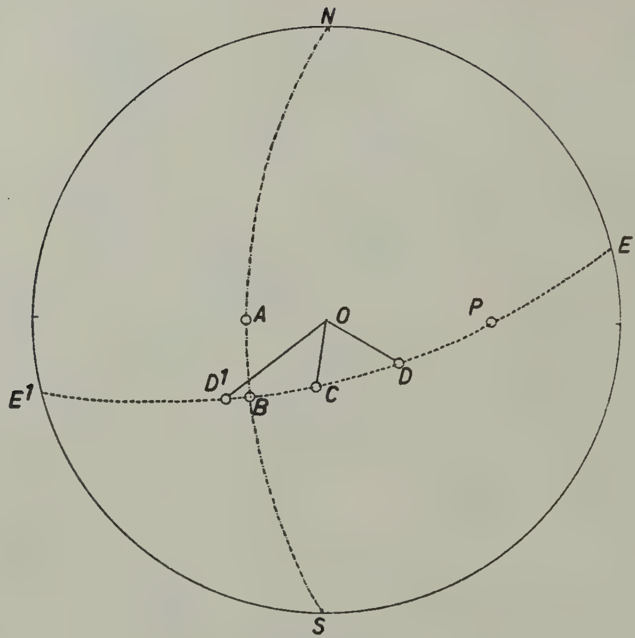
Procedure: Plot the north and south positions on the tracing paper and locate the projection of the vein as NAS (Stereogram 26). Plot the pole of the vein as P. Plot the ore-shoot as BO, plunging 50° in a direction N44°E and intersecting the vein in B.

A plane which is normal to the vein as well as containing the ore-shoot BO is defined by the great circle through B and P. Draw in this great circle noting where it intersects the circumference of the net (EE^1).

(*a*) Diamond-drilling must be accomplished in the plane EBE^1 which dips 66° in a direction N15°W.

(*b*) On the stereogram locate the direction N8°E and note where it cuts the great circle EBE^1 (in C). CO corresponds to a D.D.H. within the plane EBE^1 and it plunges 64°.

(*c*) With EE^1 coinciding with the N-S direction of the net, locate a position (D) which is 32° away from C along the great circle EBE^1. DO corresponds to the other hole which forms an angle of 32° with CO, and plunges 58° in a direction N59° 30′W. However, position D^1 on the same great circle is also located 32° away from C and since the question (*c*) does not specify, D^1O may also be a hole forming the required angle with CO. This hole will plunge 43° in a direction N51°E.



Stereogram 26

Problem 27

An inclined hole intersects a vein which strikes east-west and dips south. Owing to an initial miscalculation the hole deviated from its original course. The inclination of the hole was measured with a Tro-Pari instrument at 50-foot intervals, and averages computed for every 100 feet. The results are given as follows:

		Direction.	*Dip.*
AB	1st 100 feet	N70°W	70°
BC	2nd 100 feet	N30°W	65°
CD	3rd 100 feet	N30°E	60°
DE	4th 100 feet	N19°E	60°

Depth of hole: 400 feet.

The horizontal projection (ABCDE) of the hole and the projection of the intersections on the vein (F and G) are given in the accompanying map (Fig. 27 (*a*)).

It was decided to drill another hole in a direction AK (A being the collar). This direction is essentially parallel to the line connecting the hanging wall and foot-wall intersections (FG).

Determine:

(*a*) The elevation of the point on the hanging wall of the vein, of which F is the horizontal projection.

(*b*) The same for point G.

(*c*) The inclination of the second hole so as to intersect the hanging wall at a point having the same elevation as in (*a*).

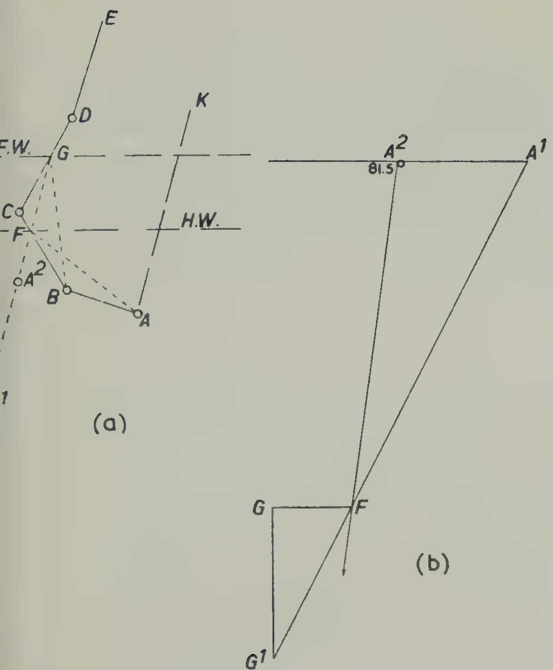
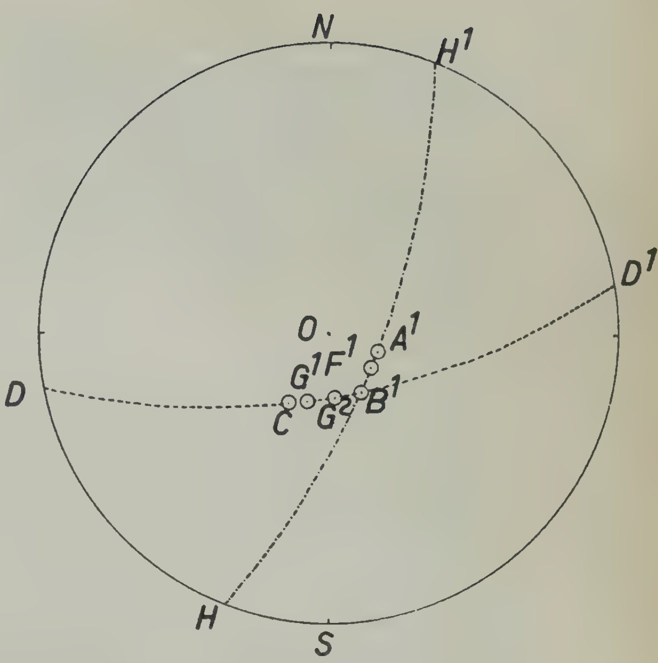


Figure 27



Stereogram 27

Procedure: For the present purpose the last 100 feet of the hole may be ignored. On tracing paper plot the first three readings on the D.D.H. as the dip vectors A^1O (AB), B^1O (BC) and C^1O (CD). If we regard the attitude of the D.D.H. in space between the two positions A and C, as two lines intersecting in B, then a plane can be constructed in which the D.D.H. will appear on the stereogram. Such a plane is given by the great circle (HH^1) through A^1 and B^1. Connect AF on the map and transfer this direction to stereogram 27. This direction intersects the great circle $HA^1B^1H^1$ in F^1. F^1O plunges 69°. Thus a line connecting the collar of the hole (A) to the point of intersection below F on the hanging wall of the vein must plunge 69° from the horizontal. The elevation difference between the collar of the hole (A) and this point below F on the hanging wall of the vein is given as

$$\text{AF (scale)} \times \tan 69° = 62' \times 2{\cdot}6051 = 161{\cdot}5'$$

(*b*) To establish the elevation of the footwall intersection proceed in a similar way. Locate the great circle through B^1 and C^1 (DD^1); transfer the direction BG (Fig. 27 (*a*)) to the stereogram and locate the point of intersection with the great circle $DB^1C^1D^1$ (G^2). G^2O plunges 66° in a direction BG. The elevation difference between B and G is given as

$$\text{BG (scale)} \times \tan 66° = 62' \times 2{\cdot}246 = 139{\cdot}2'$$

The elevation difference between the collar (A) and the position on the hole of which B is the projection is given as

$$\sin 70° \times 100 = 94'$$

The elevation difference between A (the collar) and G is found as the sum of $139{\cdot}2'$ and $94' = 233{\cdot}2'$.

(*c*) Transfer the direction FG (map) to the stereogram and locate the point of intersection with the great circle DD^1 (G^1). G^1O represents a line in space plunging 63° in a direction FG. A simple graphical solution (Fig. 27 (*b*)) shows that this line will pierce a horizontal plane, which contains A (the collar) in a point A^1. Project A to a vertical plane through G, A^1 and F (A^2). The line A^2F plunges 81°30′ in a direction FG. Since the strike of the vein is known, its hanging wall may be sketched in as shown (Fig. 27 (*a*)). All points along this line must lie on the same elevation as a point on the vein of which F is the projection. The direction AK is parallel to FG; therefore it follows that the desired hole must plunge 81° 30′.

VI CONCLUSION

Many of the problems presented here may appear somewhat hypothetical in nature. Some may remark that this approach to geological problems is purely academic. The writer feels that this is not true. It is only the engineer who finds an opportunity of solving problems involving straight lines and homogeneous planes. The geologist encounters curving lines and warped planes, but for a practical solution these must be considered as straight. Thus when dealing with a linear ore-body, an ore-shoot, a pay-streak or a lineation, he must consider them as lines, for the sake of convenience. The solution of such problems with the S.P. method may not necessarily give the correct answer, but then no other method will attain such perfection. However, the adaptability of the method, both as a time saving device and as a geometrical tool, cannot be disputed.

The method suffers from limitations, as all other methods do. Its limit of accuracy is decided by the diameter and calibration of the net adopted. Thus when the 20 cm. Wulff net is used, an accuracy of 30 minutes ($\frac{1}{2}°$) is obtained throughout. But higher accuracy may result when gently inclined planes are dealt with. On the other hand, when steeply inclined planes are plotted, an accuracy of 30 minutes is probably the maximum to be expected, if plotting is performed neatly. Ordinary tracing paper may be used if the method is adopted while solving problems in the office. However, if the geologist desires to solve problems underground, a fine transparent waxed paper is recommended as ordinary paper tends to become soiled rather easily. On the other hand if the geologist encounters problems while employed on surface mapping, the acetate as suggested by K. E. Lowe (1946) is perhaps more effective and durable than the waxed paper. A protective layer of celluloid may be superimposed on the Wulff net which is fixed to strong cardboard or light metal, to overcome extensive soiling. The net should be carried in a thin leather or cloth case with four triangular protective flaps clipping over the thumbtack. It is desirable never to remove the thumbtack but to have it glued at the bottom of the cardboard backing, as continual removal quickly enlarges the hole.

The writer is indebted to Prof. D. L. Scholtz who focused attention on the utilisation of the upper hemisphere instead of the lower hemisphere in the application of the Stereographic Projection. Thanks are also due to Dr. W. C. Brink of the Department of Geology who showed a keen interest in the progress of the work and ultimately read the manuscript critically, and Mr. A. Ravenscroft who suggested various literary alterations.

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THE DYKE ROCKS OF CAPE ST. MARTIN

By

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(Submitted January, 1956)

ABSTRACT

The geology and petrology of a small sub-parallel swarm of acid dykes traversing the marginal facies of the Saldanha batholith is described in detail. The well exposed intrusions possess distinct chill selvages and some have granite cores. The field and laboratory textural classification in use in the Geological Department of the University of Stellenbosch is discussed and applied. A protoshear hypothesis is advanced to account for the origin of primary sheets of fluid inclusions in quartz phenocrysts. Optical properties and X-ray diffraction photographs indicate that zonal microcline-microperthite of early crystallisation displays varying degrees of triclinicity. Eight new analyses are included and the differentiation trend of the dyke magma is discussed. Evidence is advanced to show that emplacement of the dyke magma was accompanied by dilation and stoping. Proof that rhyolite magma is synonymous with granite magma is believed to be afforded by the imperceptible transition from rhyolite to granite in a single well exposed intrusion displaying conclusive evidence of liquid emplacement.

This paper was awarded the Corstorphine Medal and the first prize of the Geological Society of South Africa for the year 1955

CONTENTS

I INTRODUCTION

The controversy over the origin of granite resulted in a lively revision of petrological data: more recently the centre of interest shifted into a mineralogical field, namely, the contrasting mineralogy of extrusive and plutonic acid rocks. The occurrence of high-temperature felspars has been definitely established in the extrusive and hypabyssal rocks, which the magmatic school of thought believes to be the products of a granitic magma.

The fact that these felspars differ distinctly from the low-temperature varieties found in plutonic rocks, was immediately exploited by the transformationists in an endeavour to justify their contention that granite, unlike rhyolite, is not the product of an intrusive acid magma. In an attempt to throw more light on the problem of the non-existence of granite magmas, the detailed field relations and the petrography of a minor swarm of well exposed acid dyke rocks of Cape St. Martin invading the pre-Cambrian granite of the Saldanha batholith, were investigated at the suggestion of Prof. D. L. Scholtz.

Cape St. Martin (E. longitude 17° 55′ 10″ and S. latitude 32° 42′ 52″) is the termination of a north-westerly trending peninsula some three-quarters of a mile in length along the north coast of the Malmesbury district in the Cape Province. The peninsula which is about 340 yards wide, is fringed by a rocky shore exposing at least ten separate quartz porphyry dykes invading two texturally different varieties of granite. Detailed plane table work was conducted on a scale of 1 : 600, which proved to be most suitable for the mapping of the intertidal wave-cut bench characterised by the presence of smooth unweathered outcrops ideal for purposes of careful study in the field. The most prominent features displayed by the intrusions may be said to be the chilled selvages which they possess, the progressive, textural gradation from margin to a core of granite porphyry or granite, and the evidence of multiple intrusion.

Publications bearing directly or indirectly on these intrusions and effusives of the Western Province, date back to the first decade of the last century. A brief summary of the relevant views expressed by numerous investigators and published in periodicals and textbooks, many of which are now out of print, is therefore advisable.

It is well known that in the year 1813 Captain Basil Hall discovered the Platteklip Gorge granite-hornstone contact which was first described by J. Playfair and B. Hall (1815). The famous Sea Point contact, now protected by the Historical Monuments Commission, was first described by C. Abel (1818). Twenty-six years later Charles Darwin (1844) drew attention to and satisfactorily accounted for the structural continuity displayed by xenoliths and wall rock in this exposure.

The discovery of outcrops of similar granite elsewhere in the area between George and St. Helena Bay was due partly to the work of A. G. Bain, A. Wyley and E. J. Dunn, before the systematic mapping of parts of the Western Province was undertaken by the Geological Commission of the Cape of Good Hope. A. W. Rogers and G. S. Corstorphine (1897) were impressed by the general petrographic similarity of the eight major bodies of coarsely porphyritic biotite granite, their intrusive relation to the Malmesbury series and the abundance of xenoliths. The local banding and shearing of the granite at Clifton and Stellenbosch respectively, as well as its associa-

tion with dykes of quartz porphyry at Saldanha Bay and elsewhere, was also recorded.

In 1905 A. W. Rogers published a useful short account of the geology of the granites in which he lays stress on the original gneissic structures parallel to the stratification of the sediments enveloping the Saldanha-Darling intrusion, and on the crushing of the gneiss with the production of breccias and gritty schists at Robertson and Stellenbosch. Gneissic quartz porphyry and dykes of quartz porphyry in continuity with the main intrusion traversing the slates along their strike, was observed near Darling and Paarl respectively. In referring to the Sea Point contact he states: "The slates have been thoroughly permeated by fluid granite, and have a shredded structure with granite lying between slightly bent shreds of slate."

In their textbook "The Geology of South Africa", F. H. Hatch and G. S. Corstorphine (1909) devote a few pages to the granite intrusions. The Darling-Saldanha mass is considered to be the most extensive boss while the quartz porphyry of the French Hoek valley is regarded as an apophysis from an underlying or neighbouring granite mass.

In 1912 S. J. Shand recognised and described certain veins and five different types of inclusions in the Stellenbosch granite.

E. H. Schwarz (1913) in studying the Sea Point contact, concluded that the emplacement of the granite was effected by a dual process, namely, by forceful injection and by means of solution and deposition through the agency of water. Granite substance which was injected under pressure suffused into the slate. The removal and absorption of the slaty material, followed by further injection, led to the production of schlieric structures by segregation.

In criticising this paper, D. P. McDonald (1913) expressed the opinion that Daly's stoping hypothesis would adequately account for all the observed facts.

A. R. Walker wrote two short papers. In the earlier publication (1917) he describes the geology of the intrusive Schapenberg granite and draws attention to the local abundance of tourmaline in quartz veins trending parallel to shear zones as well as in the country rock along the contact. The textural variations displayed by the granite led him to believe that intrusions of two ages are present.

In the later publication (1929) prepared for the International Geological Congress he describes the Sea Point contact. He believes that the slates were rendered plastic owing to lit-par-lit injection and permeation by the intrusive granite, while the banding displayed by the granite near Clifton is to be attributed to the partial assimilation of a large block of sediment stoped from the roof of the batholith.

In 1926 the first edition of A. L. du Toit's "Geology of South Africa" was printed. About two pages of this volume is devoted to the granites intrusive in the Malmesbury series. The granite masses of Saldanha and Darling are believed to be part of a single compound intrusion in which a younger, medium- to fine-grained granite transitional into granite- or quartz-porphyry is present in the Saldanha area. In describing the contact at Sea Point he says: "Extraordinary is the coarsely porphyritic nature of the granite in contact with the slates and the narrow veins that penetrate them . . . Clearly, under the influence of the molten granite the fragments of slate must have become softened or even melted, and movement then took place in the composite mass".

In the year 1929 another volume entitled "The Geology of the Union of South Africa" by A. W. Rogers, *et alia*, appeared. This treatise includes a very brief description of the post-Nama granites of the Cape Province.

In a comprehensive paper D. L. Scholtz (1946) concludes that the plutons of coarsely porphyritic granite which are of pre-Cambrian age almost invariably grade

upwards into a fine-grained, chilled hood facies now in different stages of erosion. The primary structures of the plutons indicate that they are locally transgressive batholithic intrusions which have been concordantly emplaced along the cores of unsymmetrical anticlines of Malmesbury sediments under waning orogenic stress. Inclusions, ranging in size from microscopic dimensions to masses several hundred feet in length, are present in the batholiths. These xenoliths exhibit various stages of assimilation up to almost completely granitised material, and this fact together with their abundance strongly suggests that stoping played an important part in the emplacement of the batholiths. The proportion of assimilated shale is reflected by the chemical composition of the hybrid rocks, while the molecular-value curves show strikingly different trends when compared with analogous curves of the differentiation diagram. In portraying the distribution of prominent shear zones and quartz porphyry dykes, a generalised method of representation has been employed owing to the small scale of the maps. There is reason to believe that granites and quartz porphyries of two distinct but very similar ages are associated in some batholiths, but the volume of the younger intrusives is relatively small.

Three months later F. Walker and M. Mathias (1946) submitted a detailed description of the Sea Point contact to the Geological Society of London. They believe that alkaline emanations, released by the invading and already contaminated granite magma, saturated and softened the Malmesbury hornstones in the contact zone so as to produce the migmatites. After consolidation the large crystals of microcline-perthite, characterising the contact and the intrusion as a whole originated by replacement through the activity of potassium-bearing solutions.

The proposed late origin of the phenocrysts evoked considerable criticism from A. K. Wells, C. E. Tilley and R. M. Shackleton (1946) as well as from F. P. Mennell and E. Spencer (1947). A neat proof of the early origin of the potash felspar phenocrysts was afforded by S. J. Shand (1949) who showed that insets of microcline-perthite may be enveloped by broad plagioclase mantles.

C. Boocock (1950) believes in the early crystallisation of the felspar phenocrysts and confirms D. L. Scholtz's conclusions about the structure, attitude and probable mode of emplacement of the Cape Peninsula intrusion. Two principal types of xenoliths are recognised. All gradations between normal hornstone and completely granitised inclusions are observed and the chemical changes accompanying such transformations are discussed.

Reference should also be made to the work of L. T. Aldrich *et alia* (1955) in their determinations of the age of the zircon and biotite from the granite of the Cape Peninsula. Determinations based on the ratios U^{238}/Pb^{206}, U^{235}/Pb^{207}, Pb^{207}/Pb^{206} and Rb/Sr respectively returned the following values: 350, 375, 550 ± 50 and 820 ± 40 million years. In spite of the fairly good agreement the first two of the four values mentioned appear to be too low. Taking into consideration the late pre-Cambrian geological age assigned to the granitic intrusions, it would seem as if the absolute age of 550×10^6 years derived from the isotopic lead-lead ratio is of the correct order.

In reviewing the publications spread over a period of almost 150 years it is clear that *all geologists*, whether magmatists, metasomatists or transformationists are unanimous on—

(i) the abundance of xenoliths in the Cape granites,

(ii) the various stages of alteration exhibited by the sedimentary inclusions and the high degree of granitisation exhibited by some xenoliths often at considerable distance from the nearest contact,

(iii) the gradational relationship which obtains between different textural varieties of granites of the same age, and

(iv) the fluid or mobile character of the granites at the time of intrusion.

The views of the leading wet-transformationist H. H. Read (1951) is of great interest on this matter. He says: "I have to confess that I was first shocked and then relieved by what I saw at Sea Point. In the past, I have used the development of identical felspars in the Malmesbury rocks at the contact and in the adjacent granite as indicating granitisation. The exposures reveal that nothing of the kind has occurred there. The contact is a moved one, showing clearly the mechanical mixing of softened-up Malmesbury slate and viscous granite material [Scholtz (1946, plate IV) illustrates this contact]. It is true it is a mixed contact, but of a mechanical kind . . . My relief arose from the realisation that the Cape granite body was in the wrong geological setting for large scale granitisation; . . . . I suggest that the Cape Plutons, like the other granite bodies that I call plutons, made their way in as *nearly solid** masses by softening-up the country rocks round about them . . . even in the wrong setting pocket examples of granitisation could be observed. Thus, Professor Scholtz showed me, on the coast north of Groot Paternoster near Saldanha Bay, great rafts of metasomatised Malmesbury sediments felspathised and soaked and veined by granitic materials; this zone passed fairly rapidly into porphyritic granite of Cape Type, containing rounded relics of transformed sediments which gave place to homogeneous granite. From such a section much detail on granitisation can be observed and many valid conclusions may be drawn—but the conclusion that all the granite of this pluton is of *granitisation-origin at its present level* is certainly not valid—once again I have to remark that the regional setting is against it."

From the foregoing it is clear that indisputable field evidence enforces the conclusion that the granite body was mobile at the time of its emplacement and that in the case of the Cape granites the much flaunted "room problem" has lost its argumentative significance.

Now the granitisation of xenoliths and down-stoped rafts of sedimentary rock several hundred feet in diameter when immersed in a magma chamber of batholithic dimensions, presents no obstacle to the magmatist. Neither does the local metasomatism or granitisation of the wall rock introduce any insuperable difficulties, provided that the transformation occurs on a scale commensurate with the size of the intrusion. Since the granitising ability of a granite magma is a function of its volume and the depth of emplacement, it follows that the magmatist is not faced with the problem of stretching his imagination to the utmost limits by postulating granitisation on a regional scale.

It is only too clear that granitisation on a small or proportional scale as defined above is a fundamentally different concept to wholesale granitisation without the intervening agency of a magmatic medium. The fundamental distinction between the magmatic and transformationist schools of thought lies respectively in the *mobile or solid state* of a given body of granite in the process of evolution and emplacement. Any irrefutable evidence of the original mobility of a granite mass is, therefore, the hallmark of magma, and any transformationist who seeks to evade the issue by transferring the seat of granitisation to the inaccessible depths of the earth's crust is merely labouring under an illusion.

The writer believes that this brief summary of the current views on the so-called Cape granites will serve as a background for the present investigation, in which other aspects of the granite problem will be discussed.

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* Italics of the author of this paper.

II GENERAL GEOLOGY AND STRUCTURE OF THE GRANITES

The rocks underlying the Cape St. Martin peninsula comprise part of the northern rim of the Saldanha batholith, but the nearest known granite-hornstone contact is exposed along the shore about two miles to the east of the area under consideration. Apart from a few isolated exposures of granite, dunes of wind-blown sand admixed with varying amounts of fragmental shelly material cover the central and major portion of the peninsula. Along the shore, outcrops of granite are intermittently interrupted by local, often thick, accumulations of boulders and shingle consisting of quartz porphyry, different textural varieties of granite and hybrid rocks.

When one follows the rocky intertidal portion of the shore around the peninsula, twenty independent outcrops of quartz porphyry dykes are to be seen. Owing to the bifurcating and tortuous habit, as well as the accompanying textural variation along the strike of these intrusions it is not possible to reconstruct the solid geology of the peninsula without error. A fair estimate of the number of dykes within the surveyed area would be ten, but many others are known to occur beyond its borders.

The granites invaded by the quartz porphyries not only vary in texture but also in mineralogical composition and colour index. The fine- to medium-grained granite, which is the most abundant variety in this area, can be correlated with the earlier batholithic hood facies, while the coarse porphyritic granite, which underlies most of the area beyond the limits of the map, comprises part of the core of the Saldanha Bay batholith.

The fine-grained granite of the peninsula locally grades into a medium-grained variety in which the essential minerals may attain diameters of 4 mm. The potash felspar is microcline, showing the usual perthitic segregations, but the quadrille twinning is notably absent. Microscopically the plagioclase occurs in subhedral grains, which vary in composition from albite to oligoclase with occasional zoning and saussuritisation of the cores. Up to 5 per cent of biotite, in small irregular crystals or haphazardly orientated clusters, may be present in the rocks. Accessory minerals are magnetite, zircon and skeletal crystals of allanite almost completely altered to a secondary yellowish isotropic product, which is responsible for the staining of the enveloping minerals.

In chemical composition the even-grained granite closely resembles the chilled, fine-grained hood facies of the Malmesbury batholith taken near the summit of Paardeberg.

	No. of Anal.	*si*	*al*	*fm*	*c*	*alk*	*k*
Even-grained granite, Cape St. Martin	1	480	47·2	8·3	7·2	37·3	0·43
Fine-grained hood facies, Paardeberg	1	466	44·9	8·1	7·4	39·7	0·49
Outer 20′ chilled zone, Slippers Bay	3	491	47·9	8·5	3·5	40·1	0·47
Average composition 1,200′ chilled zone, Slippers Bay	6	461	44·5	11·3	4·8	39·4	0·34

About eight miles south-east of Cape St. Martin a non-porphyritic fine- to coarse-grained marginal chill facies, approximately 1,200 feet wide, occurs in contact with the hornstone at Slippers Bay. This area is now being investigated in detail by J. R. McIver.

From six of his new analyses of rocks from the marginal facies of the Saldanha batholith, average Niggli values of the outer and finest grained 20-feet-thick contact zone, as well as of the entire zone, have been calculated.

From the listed molecular values it is clear that, with the exception of *alk*, the Niggli values of the three undoubted marginal rock types differ more greatly from one another than any one of them does from the values of the Cape St. Martin granite.

The porphyritic granite, somewhat less coarse than the coarsely porphyritic granite of Cape Town, forms the massive outcrops seen at the base of the peninsula, but two other small exposures also occur along its north-western coast.

The rock is moderately coarse-grained with phenocrysts some 3 cm. in length set in a matrix whose grain size ranges up to 5 mm. Texturally this granite grades into finer non-porphyritic types on the north-eastern part of the coast, just beyond the area mapped.

The porphyritic granite contains an abundance of hornstone inclusions in various stages of assimilation and incorporation. The inclusions may possess knife-sharp margins or may be characterised by the presence of a diffuse rim separating the inclusion from the granite (Plate II, phs. 1, 4, 6). Small biotite-rich inclusions are often arranged parallel to the lineation of the granite and, in general, xenoliths are abundant right up to its contact with the fine-grained granite.

The contact between the two granites is exposed in a few places on the northern side of the coast. The southernmost contact exposed on the north-eastern shore is sharp and is characterised by angular re-entrants, and large phenocrysts of potash felspar are present in the porphyritic granite up to its junction with the even-grained granite. The fine-grained granite at this contact is much the same as elsewhere on the peninsula, but along its north-eastern coast the fine-grained granite shows a notable change in character. The lineation becomes indistinct, faint schlieric segregations of biotite develop and phenocrysts of potash felspar appear and become more numerous as a small exposure or window of coarsely porphyritic granite in the overlying even-grained granite is approached. In the immediate vicinity of the northernmost out-crop of porphyritic granite equant insets of the potash felspar again appear in the surrounding even-grained granite within fifty feet of the indistinct contact.

The same phenomenon also characterises the junction between the two granites on the opposite shore of the peninsula, and in the immediate vicinity of all three contacts the usually smoothly-weathered surface of the fine granite is seen to be distinctly scabrous. It is noteworthy that no aplites were observed to cross any of the contacts.

Since these peculiarities appear to be confined to the rocks in the immediate vicinity of visible contacts, it is inferred that the porphyritic granite is relatively close to the present surface on the north-eastern side of the peninsula, where these phenomena characterise the even-grained granite over a considerable area.

Aplites appear to be equally abundant in both varieties of granite. They are exceedingly fine-grained and may show preferred orientation of biotite flakes parallel to their walls. The veins and dykes which dip at high angles range in width from less than an inch to slightly more than a foot. Aplites associated with this even-grained granite are brown in colour, which serves to distinguish them from the pale grey dykes and segregations in the porphyritic granite. Irregular or sheet-like segregations of aplite, some dipping at low angles to the horizon, are relatively uncommon. In one direction they have been observed to grade into pegmatite or porphyritic granite, whereas in the other the borders are fine-grained and fairly sharply defined.

In general, lineation is not readily visible throughout the porphyritic granite but in isolated areas it is well developed and shows a common direction of orientation. In such localised areas the stretching may be deduced from the preferred orientation of any of three of the four commonest minerals. For example, the north-westerly strike and sub-horizontal attitude of the lineation in the porphyritic granite are revealed respectively by the quartz lenticles and plagioclase laths on the north-eastern and south-western shores of the peninsula, while the attitude of both the lineation and the foliation may be deduced from the orientation of the biotite flakes. In the even-grained granite the lineation plunges at an angle of 25 degrees in a northerly direction whereas its strike, which is north-north-west, is inclined to the trend of lineation of the porphyritic granite at an angle of approximately 18 degrees.

In an attempt to elucidate the structure of the area and its relation to the attitude of the dyke intrusions, the direction of strike and dip of more than a thousand joints as well as that of the shear planes were measured and plotted on a Schmidt equal-area net. For convenience the peninsula has been subdivided into three structural areas, A, B and C, in each of which the characteristic structural pattern of the granite and its associated aplites is indicated by means of composite statistical diagrams. The joint system characterising the dyke intrusions appears in a separate but also geographically orientated selective diagram (Fig. 1D).

From the composite statistical diagram, Fig. 1C, it will be observed that the maximum of the most prominent set of joints is at right angles to the direction of the lineation in the porphyritic granite. By definition they are accordingly the q-joints which strike in a north-easterly direction. The group of quartz porphyry dykes invading the porphyritic granite of the southern half of the peninsula also follow this direction. It is noteworthy that the half-dozen aplites seen in this region do not appear along q-joints but strike rudely parallel to two intersecting planes which form an obtuse angle with the direction of stretching.

In the B area (Fig. 1) the invaded granite consists almost wholly of the older even-grained marginal variety in contact with the porphyritic granite of the central type.

It will be seen that the q-joint maxima show a slight easterly change in direction and this variation is further accentuated near the apex of the peninsula in the area A. Steeper dips and greater fluctuation in the direction of strike are also apparent when one goes in a northerly direction. This variation and the fact that the plunge of the lineation is not strictly normal to the q-joint surfaces, as in area C, may be ascribed to the influence of the underlying porphyritic granite, which is probably close to the surface.

Highly inclined joints almost parallel to the strike of the lineation appear to be best developed in the porphyritic granite (Fig. 1C). They are probably primary since they are exploited by about six aplite veins as well as several minor, broad but sharp, protuberances from a few of the later intrusions comprising the northern group of quartz porphyry dykes in area A. This set of steeply dipping joints almost normal to the q-joints, may therefore be regarded as belonging to the s-joint system, whose maxima in the composite diagrams for the areas C and B is seen to migrate from the north-east to a true easterly position. In the area A further to the north-west these joints are probably tight and obscure except where exploited by wave erosion. The strike of the s-joints therefore not only exhibits clockwise rotation, as in the case of the q-joint system, but simultaneously also shows a change in the direction of the steep dip from west to east.

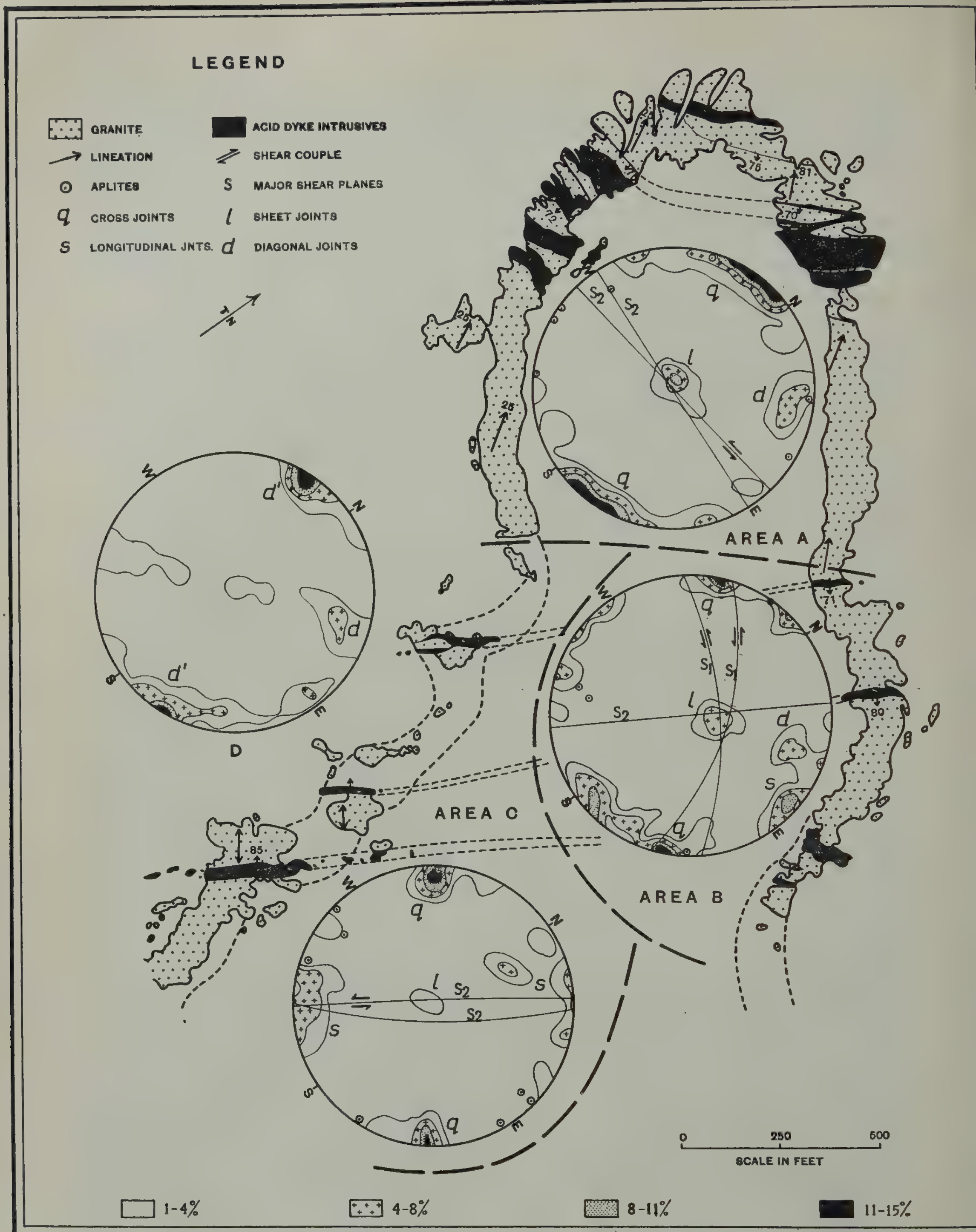

FIG. 1. Composite geographical diagrams of aplites, shear planes and approximately 200 granite joints in each of the areas A, B and C. D is a selective diagram of 340 joints in the dyke intrusions.

Sub-horizontal jointing, though fairly well developed in the even-grained granite (Fig. 1A) is not easily recognisable in the porphyritic granite. The absence of aplites along these joints, together with the fact that they also extend into the dykes, seems to indicate that they are not true l-joints but flat-lying sheeting of secondary origin.

While the area was being mapped, another minor set of steeper westerly-dipping joints trending in a north-west—south-east direction was observed to traverse both intrusive and country rock. It is possible that these joints may be correlated with the late d-joint system which are known to occur in other granite massifs subjected to continued orogenic stress after consolidation.

The granites of the peninsula are highly sheared in certain directions and well-defined shear planes vary from a fraction of an inch to two inches in width, and the feather shearing about these shear planes shows the same direction of displacement as in the master shear, as pointed out by H. E. McKinstry (1953). In the north-eastern part of the coast (Fig. 1B) the north-west—south-east trending directions of shear are older than the dykes, and the movement along them results from a counter-clockwise couple.

These S_1 shear planes, which are characterised by impersistent films of mylonite and a good deal of chlorite, must be very old since they appear to have been exploited by impersistent veins of material resembling aplite. The five-hundred-feet-broad zone of highly sheared rocks, developed in the vicinity of the coarsely porphyritic granite-hornstone contact in North-West Bay, originated in the interval which elapsed between the consolidation of the intrusive granite and the injection of its aplitic differentiate [Scholtz, 1946 (pp. lxv and Plate V, Fig. 2 and Plate XVII, Fig. 2)] and before the emplacement of the younger even-grained granite exposed south of Cape Castle, corresponds closely with the S_1 shear zones of Cape St. Martin in age, trend and anticlockwise couple orientation.

The next most important epoch of counter-clockwise shearing which resulted in the production of the S_2 shear zones (which cut both dykes and invaded rocks of the peninsula) is the northern equivalent of the less conspicuous north-east—south-west shear zones traversing both older and younger granites south of Cape Castle (op. cit. Plate XVII, Fig. 1 and Plate XVIII, Fig. 2). A characteristic feature of the shearing in both regions is the low order of the displacement. Nicks or serrated indentations of the dyke walls (Plate I, dykes *d, e, f*), which on casual inspection may be regarded as evidence of an intrusive relationship, in reality owe their origin to such minor displacements along one or several closely spaced S_2 planes. In area A (Fig. 1) it will be seen that the poles of S_2 planes, like the joints and lineation in the even-grained granite, have migrated eastwards relative to their orientation in the porphyritic granite.

It should also be mentioned that evidence of later epochs of feeble shearing and crushing of the granites and dyke rocks is also present. Minor shears offsetting the S_1 and S_2 types in a clockwise direction are usually accompanied by quartz veining and epidotisation.

Seven exposures of much-fractured rock belts, 25-100 feet broad, are to be seen along the shore. If the relevant boundaries of exposures on opposite shores of the peninsula are connected, the existence of two continuous south-west—north-east trending fracture zones is suggested. These belts of fractured rock represent the products of the youngest of all earth movements in the area under consideration.

III FIELD PETROLOGY OF THE DYKES

A glance at the geological map (Plate I) shows that the dykes of the northern group which invade the even-grained granite near the termination of the peninsula are more irregular and arcuate than those comprising the southern group which penetrate the less disturbed porphyritic granite. It will be observed that, unlike the dykes of the southern group, the opposite extremities of some of the intrusions of the northern group cannot be positively correlated. For this reason as well as to facilitate description, a letter has been assigned to each of the twenty outcrops occurring along the shore of the peninsula. In this chapter an attempt will be made to describe the megascopically visible textures, structures, form and characteristics of the dykes, as well as the relations between these intrusions and the structure of the country rock.

In areas B and C (Fig. 1) the strike and attitude of the dykes of the southern group display a marked parallelism to the q-joint pattern of the invaded rocks. A consideration of the equatorial spread of the q-joint maxima in the statistical diagram A and the arcuate trend of the dykes dipping 70 degrees to 90 degrees south-east in the northern area corroborates this conclusion in a still more striking manner. In the same figure the selective-joint diagram D represents all the dyke joints of the mapped area. In this geographically orientated diagram the maximum is mainly attributed to the well jointed dykes of the southern area. Since the trend of the dykes makes an angle of 40-45 degrees with this most prominent system of fractures, a diagonal type of joint appears to be indicated. R. Balk (1937, p. 38) ascribes the origin of such joints to the welding of fine-grained rocks on to coarser country rock in the case of dykes; "fine-grained 'interpositions' commonly develop diagonal joints even though the surrounding country rock may fail to show them". The d-joints are best developed in the parallel walled dykes invading the porphyritic granite. Joints parallel to the walls but within the intrusions are poorly represented in the field. This fact, when considered together with the variable strike of the dykes, adequately accounts for their scattered and weak representation on the statistical diagram D. It will be recalled that the youngest joint system though not particularly well developed, tends to maintain a relatively constant orientation in the granite. That the poles of these joints on the diagrams A, B and D (Fig. 1) indicate relatively slight migration and that they not only traverse the invaded granite but also the dykes irrespective of their orientation, seem to indicate that these fractures are geologically younger than the dyke joints previously described.

The relative importance and general inter-relations of the joints characterising the granite basement of the peninsula are revealed by the collective statistical diagram, Fig. 2. The d- and s-joint fields partially overlap and are less prominent than the fairly well defined sub-horizontal sheeting fractures. The elongation of the density contours and the steep south-easterly dip of the best developed q-joint system harmonise with the fluctuations in the strike and the prevalent direction of dip of the dyke swarm.

In contrast to the dykes of the southern area, the curvilinear habit of the intrusions of the northern group appears to be related to a pronounced multiformity of the dyke walls. On the geological map (Plate I) dyke *d* of the northern area is seen to show pinching and swelling, while the smooth and almost parallel walls of dyke *e* contrast markedly with the irregular boundaries of intrusion *f*. The adjacent dyke *g* with relatively straight borders maintains a more or less constant width of seven feet, while the 75-feet-wide, rather irregular, sinuously-walled intrusion locally displays the rather uncommon angular type of contact (Plate II, ph. 7). Dyke *i* which has

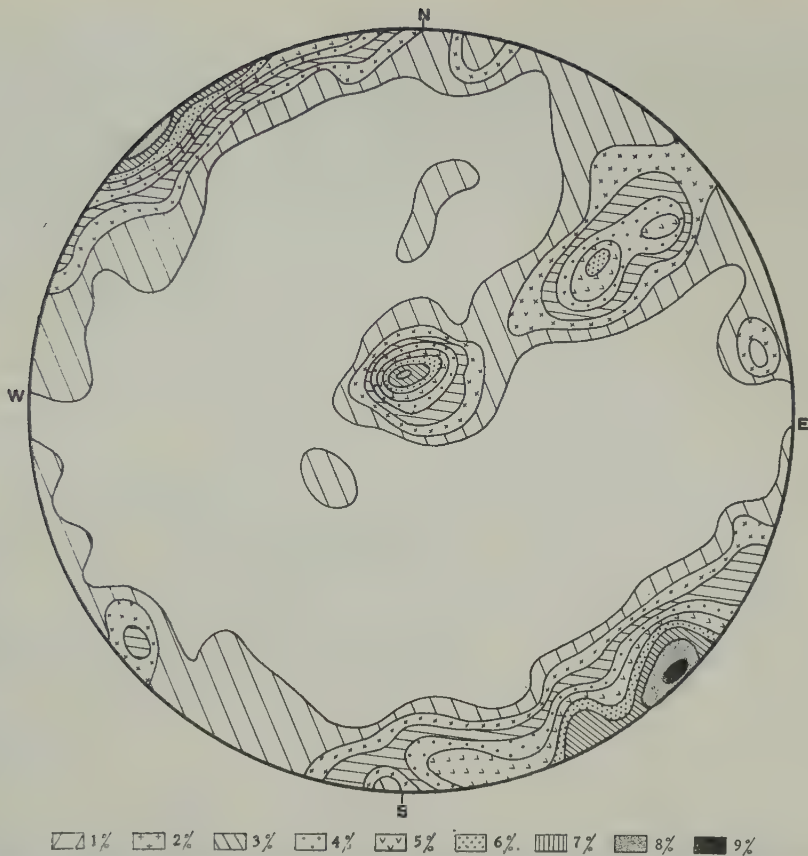

FIG. 2. Collective geographical diagram representing 650 joints in the granite country rock of the Cape St. Martin peninsula.

relatively straight but smooth borders, and varies in width from 5 feet to 10 feet, is followed by the slightly arcuated dyke *j*, whose walls are sub-parallel but undulating. Dykelet *k* represented by the stippled line on Plate I and exposed within the inter-tidal portion of the coast is remarkable in that it can be followed in a westerly direction for a distance of at least 350 feet (Plate III, ph. 7). At the point *k* it is only one and a half feet wide. This width is maintained for the first 110 feet, after which a local widening to two feet is apparent over a distance of some ten feet along its length. Beyond this local dilatation it is seen to wedge out gradually along the invaded q-joint plane narrowing down to about two inches in the extreme west.

The V-shaped dyke *l* whose northern arm attenuates in a south-westerly direction displays a distinct intrusive cusp along an s-joint plane of the invaded equigranular granite, while the southern arm, which appears to widen westwards, is intersected by a north-westerly trending marine corrasion furrow infilled with boulders. It is possible that this debris conceals the junction of dyke *l* with the massive multiple intrusion *m*, in which case it should be regarded as an apophysis. A thin wedge-like screen of equigranular granite also appears to separate the dyke *n* from the smooth, though undulatory-walled, multiple intrusion of which *o* is probably also an apophysal dykelet.

The dykes *pc*, *qb* and *ra* of the southern area, vary in width from twenty to more than forty-five feet and probably exceed 1,200, 1,500 and 1,800 feet in length respectively. The two first named have fairly regular and parallel walls and both are characterised by the presence of protuberances of country rock (Plate I, outcrop *c* and Fig. 3). The last named intrusion, which has fairly straight boundaries on the south-western coast, traverses the contact between the two granites and also the area of prominent pre-dyke shearing on the north-eastern coast, where the dyke shows appreciable irregularities along its walls. The pre-dyke features in the country rock

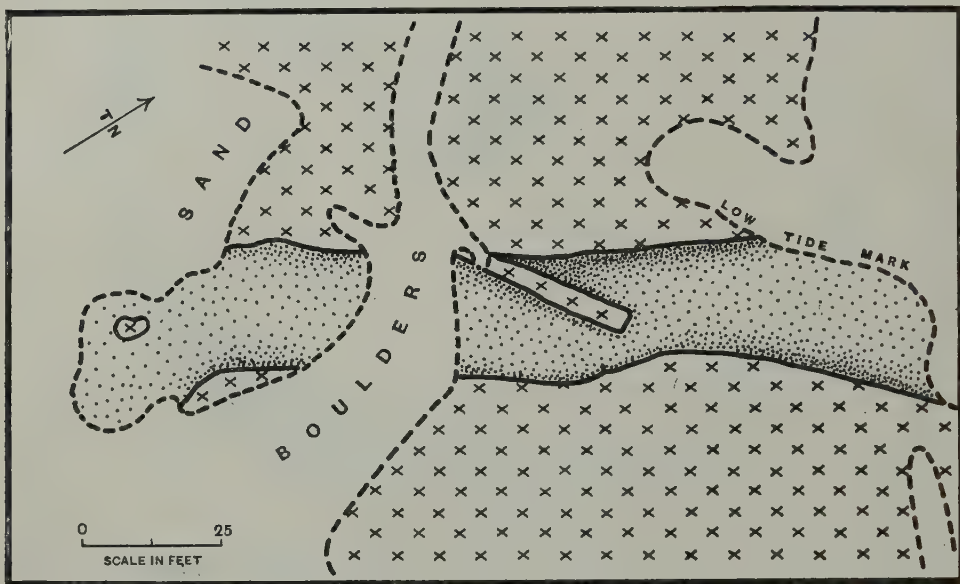

FIG. 3. Dyke *q* with chilled selvages.

such as aplite veins, shearing, etc., at this locality as well as the configuration of opposite walls do not match. Adjacent to this dyke an irregular pipe-like body dipping north-east occurs, abutting against an aplite vein which shows no offset. Except for these occurrences no other contacts suitable for the determination of offsetting were encountered along the shore of the peninsula.

The intrusive contacts appearing on the map show the true form of the dykes unaffected by the later anti-clockwise shearing movements, with a few minor exceptions involving local displacements not exceeding five feet in the case of dykes. Analogous post-dyke shear planes are also observed to traverse the massive and extensive outcrops of quartz porphyry of the Saldanha Bay area and elsewhere. These intrusives and contemporaneous effusives cut across and include the porphyritic granite of the Saldanha batholith in different localities over a minimum distance of some 25 miles from Massenberg near Langebaan to North-West Bay in the north. Along the coast about three miles north of the last named locality the same porphyritic granite is also invaded by the younger Cape Castle granite already referred to in Chapter II. Further proof that these quartz porphyries, the younger granite and the dykes of the Cape St.

Martin and Stompneus Bay areas are the same age is borne out by the crosscutting relation they bear to the shear zones of the Saldanha batholithic granite and the later but milder epoch of shearing which affected all the younger intrusives as well (Scholtz, 1946). Though all the dykes of the Cape St. Martin peninsula were emplaced between the S_1 and S_2 epochs of shearing (Fig. 1), minor and local variations in the relative ages of the intrusions can also be deduced from a study of chilled selvages and transgressive apophyses.

Though no vein-like offshoots were observed in the intrusions of the southern group, the dykes of the northern area are characterised by no less than fifteen dykelets and veins varying from a fraction of an inch up to approximately five inches in width. Apart from these, some 12 apophysal protuberances and sharp terminating projections were encountered, varying from a few inches to ten feet in width, which jut out at right angles from the dyke margins exploiting the s-joint direction of the granite. Seven of the vein-like offshoots follow this direction while the rest, which are inclined to the dyke contacts, constitute apophyses which may merge into dykelets or veins that strike approximately parallel to the trend of the parent intrusion when its contact is uneven (Plate III, ph. 1). They may either penetrate the country rock irregularly for a few inches (for example Plate III, ph. 5) or follow steep dipping, well defined fractures or joint planes for distances ranging from a fraction of a foot to several feet from the contact. On the southern border of dyke *i* an offshoot is seen to turn sharply and follow the strike of the contact for approximately 25 feet before re-entering the parent intrusion. Usually the colour and texture of an apophysis closely resemble those of the chill selvage of the parent intrusion and occasional phenocrysts of plagioclase, microcline and quartz may be visible in an aphanitic groundmass. Irregular dykelets of a somewhat coarser texture, varying in width from a quarter of an inch to six inches, occur in the country rock between dykes *f* and *h*. All these minor intrusions are cut in turn by dyke *g* which differs from them and other dykes in colour, texture and chilling.

Of all the internal characteristics the chilled selvages of these dykes are the most striking features to be observed in the field. Variations in the degree of chilling are not only visible from dyke to dyke but also appear along the contacts of a single intrusion. In some instances the chilled selvage may double its width within ten feet along the same wall whereas in other places it maintains a regular width, particularly in the narrow parallel walled dykes. Vertical sections of more than four feet are rarely visible in the field so that the amount of chilling, so to speak, can only be studied on a horizontal plane, an observation which is valid for all the dyke outcrops on the peninsula. The northern group of dykes, already characterised by their close spacing and irregular contacts, also show the greatest variation in chilling from dyke to dyke.

The density of the stippling in Fig. 4 serves to illustrate the nature and degree of chilling exhibited by six of the dykes of the northern area. Texturally the matrices of dykes *e* (Plate III, ph. 2) and *g* are relatively homogeneous save for differences in crystallinity which appears to be a function of the relative widths of the two intrusions. Attention should, however, also be drawn to the fact that *g* is one of the youngest intrusions in the area under consideration. In dykes *i* and *f* chilling effects which appear to be restricted to the narrower portions of the intrusions are respectively symmetrical and unsymmetrical, while in dyke *d* the chilling effects are accentuated and gradational when followed from the northern contact to the core of this intrusion. Dyke *h* is a multiple intrusion, in which the borders of the younger components may either be partly gradational and therefore indistinct, or sharply defined and readily recognisable owing to differences in texture. In this case the effects of chilling appear to be restricted

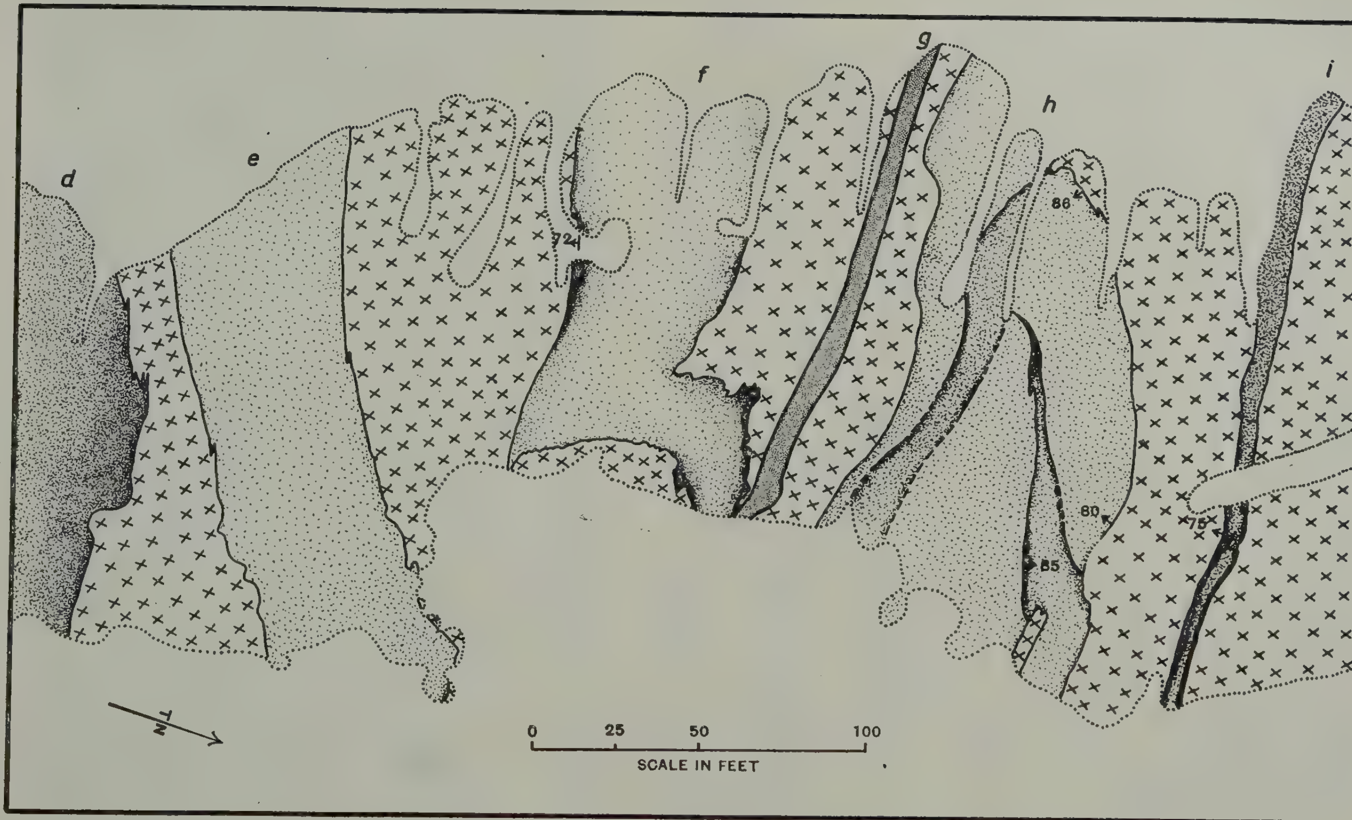

FIG. 4. Geological map in which the differential chilling (diagrammatically represented by the density of the stippling) of some of the Cape St. Martin dykes is illustrated.

to the margins of the later components, whose cores closely resemble the rock comprising the parent body in colour, texture and type of phenocrysts. It is noteworthy that material of an aplitic character, as well as coarser dyke material in the form of stringers striking north-south and dipping at 55 to 85 degrees, and a vein three inches in diameter dipping 70 degrees west, traverse the greater part of the multiple dyke *h* and cross the southern border of the intrusion to disappear into the enveloping granite in the form of wedges within a foot or two of the contact. The vein, whose average grain size is 0·3 to 0·2 mm., occasionally holds megascopically visible clusters of felspar and quartz. A few one-inch veinlets, though much finer in texture and similar to the chilled apophyses previously described, were also observed in dyke *f*. Microscopically the microcrystalline fabric and scales of biotite are seen to be orientated parallel to their well defined margins, which suggests the exploitation of fractures in the parent body. The multiple character of the intrusion *m* on the opposite shore of the peninsula strongly suggests that it is the counterpart of dyke *h*, in which the northern younger component still retains its identity. The northern chilled margin of the later intrusion which dips south-east at 81 degrees to 85 degrees is distinct and almost parallel to the northern contact of the parent dyke, while the southern margin of the younger intrusion is not only impersistent but also indefinite. This partial gradational relationship between the two components of different age, as well as the fact that aligned phenocrysts are observed to cross the inferred continuation of the southern contact of the later intrusive in the vicinity of the low tide mark, strongly suggests that the parent dyke had not completely congealed at the time of invasion.

Considered as a whole, dyke *j* shows no marked difference in groundmass texture at its north-western end, but when it was followed in an easterly direction the effects of marginal chilling become apparent.

The extremities and margins of the wedges of dyke material along the northern walls of the dykes *pc* and *qb* along the west and east coasts respectively are characterised by distinctive borders which are narrower, however, than the chilled selvages of the intrusions. Prominent salients and re-entrants along dyke walls usually display distinctive differential chilling effects, and the fact that the broader selvages border embayments in the country rock can be attributed to the more rapid cooling of the intrusive during its emplacement. This phenomenon, though not uncommon, is particularly well displayed by dyke *ra* whose dark flinty selvages are distinctive when compared with the borders of an intrusive such as dyke *e*, in which the effects of marginal chilling are barely recognisable (Plate III, phs. 3 and 6). In the selvage of dyke *ra* occasional pale streaks parallel to the walls of the intrusion and occasional slight drag effects suggestive of flow, were observed.

A patchy preferred orientation of felspar phenocrysts parallel to the borders of an intrusion or about xenoliths (Plate II, ph. 2) may occasionally be seen but, in general, prominent flow structures are rarely encountered in the dykes of the peninsula. In some instances the long axes of phenocrysts strike obliquely to the margins of an intrusion and in dyke *d* such a north-south alignment of the felspars may be seen right up to its contacts with the country rock.

Inclusions are relatively unimportant in the intrusions of the peninsula. Many dykes are barren while others may contain a few xenoliths closely resembling the country rock. In some dykes wedge-like marginally chilled apophyses of the intrusive may be seen to separate partly detached angular blocks whose form may match the configuration of the adjoining wall rock. Xenoliths of granitic material are usually angular or sub-angular, with or without narrow indistinct borders, but inclusions of hybrid granite or granitised Malmesbury sediments are usually rounded or irregular

Matrix		Grain size	AVERAGE LENGTH OF PHENOCRYSTS (LABORATORY AND FIELD)							
			> 5% PHENOCRYSTS			< 5% PHENOCRYSTS				
			> 3 cms	3–1 cms	< 1 cm	3–1 cms	1.0–0.1 cm	< 0.1 cm		FIELD CLASSIFICATION BASED ON MATRIX APPEARANCE
X EUCRYSTALLINE	PHANEROCRYSTALLINE	> 3 cms	COARSE TO GIANT (*GRANITE*) PEGMATITE						1	PEGMATITIC
X EUCRYSTALLINE	PHANEROCRYSTALLINE	1.0 – 3.0 cms	PORPHYRITIC (*GRANITE*) PEGMATITE	SERIATE (*GRANITE*) PEGMATITE					2	PEGMATITIC
X EUCRYSTALLINE	PHANEROCRYSTALLINE	0.5 – 1.0 cm	VERY COARSE PORPHYRITIC PEGMATITIC (*GRANITE*)	COARSE PORPHYRITIC PEGMATITIC (*GRANITE*)	PEGMATITIC (*GRANITE*)				3	PEGMATITIC
X EUCRYSTALLINE	PHANEROCRYSTALLINE	1.0 – 5.0 mm	VERY COARSE PORPHYRITIC (*GRANITE*)	COARSE PORPHYRITIC (*GRANITE*)	PORPHYRITIC (*GRANITE*)	COARSE TO MEDIUM EVENGRAINED (*GRANITE*)			4	GRANITIC
D DYSCRYSTALLINE	MICROCRYST.	0.1 – 1.0 mm	VERY COARSE (*GRANITE*) PORPHYRY	COARSE (*GRANITE*) PORPHYRY	(*GRANITE*) PORPHYRY	COARSE PORPHYRITIC MICRO (*GRANITE*)	PORPHYRITIC MICRO (*GRANITE*)	MICRO (*GRANITE*)	5	APLITIC
D DYSCRYSTALLINE	CRYPTOCRYST.	0.001–0.1 mm	VERY COARSE (*RHYOLITE*) PORPHYRY (=Very Coarse Quartz Porphyry)	COARSE (*RHYOLITE*) PORPHYRY (=Coarse Quartz Porphyry)	(*RHYOLITE*) PORPHYRY (=Quartz Porphyry)	COARSE PORPHYRITIC (*RHYOLITE*) (=Coarse Porphyritic Quartz Porphyry)	PORPHYRITIC (*RHYOLITE*) (=Quartz Porphyry)	(*RHYOLITE*) (=Liparite)	6	FLINTY
D DYSCRYSTALLINE	HOLOHYALINE	holohyaline or devitrified	VERY COARSE (*PITCHSTONE*) PORPHYRY VERY COARSE (*VITRO*) PHYRE	COARSE (*PITCHSTONE*) PORPHYRY COARSE (*VITRO*)PHYRE	(*PITCHSTONE*) PORPHYRY (*VITRO*) PHYRE	COARSE PORPHYRITIC (*PITCHSTONE*) COARSE PORPHYRITIC (*OBSIDIAN*)	PORPHYRITIC (*PITCHSTONE*) PORPHYRITIC (*OBSIDIAN*)	(*PITCHSTONE*) (*OBSIDIAN*)	7	GLASSY
			I	II	III	IV	V	VI		

FIG. 5.

and streaky and tend to merge imperceptibly with the intrusive. All along the coast between Cape St. Martin and Klein Paternoster, some six miles to the south-south-west, inclusions of hornstone and hybrid microgranite in all stages of assimilation are abundant in the Saldanha granite, and many xenoliths of similar material were also observed to occur in a broad coarse intrusive dyke 1,300 feet from Cape St. Martin, which when it is followed in a south-westerly direction is seen to become more contaminated owing to an increase in the number of inclusions, which are probably derived from the adjacent mass of hybrid granite. On the peninsula, however, only one well rounded inclusion of microgranite surrounded by a transitional or mixed border was found in dyke *a*. The mixed material extending into the central portion of this xenolith, illustrated in Plate II, ph. 5, also contains potash felspar phenocrysts identical to those of the intrusive, as well as bipyramidal β-quartz crystals which stand in relief on the weathered surface.

IV PETROGRAPHY

A. Texture and Mineralogy of the Dykes

According to P. Niggli (1954) and A. Johannsen (1941) continental geologists apply the term quartz porphyry to acid volcanites or the pre-Tertiary extrusive equivalents of granite in order to distinguish such rocks from the geologically younger lavas of similar composition. The fact that the acid Mesozoic lavas of the Lebombo range are termed rhyolites, while the acid extrusives of Ventersdorp age in the south-western Transvaal are described as quartz porphyries, clearly indicates that this distinction has not been recognised in South Africa.

Since some writers also regard quartz porphyry as the hypabyssal rather than the volcanic equivalent of granite, while other petrologists have shown that it may constitute a local chill facies of stocks and batholiths of granite, it would seem advisable to regard the term quartz porphyry as being synonymous with rhyolite, as suggested by writers of several modern petrological textbooks.

Because no purely descriptive textural classification of the eruptive rocks has yet been published, the writer has included the laboratory and field classification used in the Department of Geology of this University. Recognition of the great variation in the texture of the Western Province granites and associated igneous rocks and the petrological significance of the variation in the texture led to the formulation of a uniform and simple scheme of textural classification (Fig. 5) which could be used by research workers conducting detailed investigations in different localities. In the case of the above-mentioned rocks of the granite-rhyolite clan, a coarsely porphyritic granite and a porphyritic rhyolite the equivalent of quartz porphyry, may be distinguished texturally by the symbols X.II.4 and D.V.6 respectively. Though the subdivisions are partly based on the sizes of phenocrysts observed in the eruptive rocks of the Western Province, it is clear that the scheme of classification may be equally satisfactorily used to systematise textural variation in other rock clans (cf. foyaite-phonolite, syenite-trachyte and so on) and should therefore prove useful when used together with the Shand Symbol.

Dyke *f* was selected for special study on account of its well developed granitic core which is exposed near the low-tide mark. In handspecimen the chill phase of this intrusion has a bluish grey, flinty groundmass holding more than 5 per cent of recognisable phenocrysts whose average length is slightly less than 1 cm. This rock can therefore be described as a rhyolite porphyry (quartz porphyry) with symbol D.III.6.

The microscopic study of samples 1 to 3 (Plate IV, pms. A, 1 to 4) of the chill phase reveals a gradual increase in the grain size of the matrix from 0·01 to 0·07 mm. and two generations of phenocrysts, of which the insets of the second generation are seen to increase progressively in abundance towards the interior of the intrusion. Though most of the dykes of the peninsula fall into this textural group the principal local variation observed is to the textural type D.II.6 or coarse rhyolite porphyry owing to the greater abundance of large phenocrysts (Plate III, ph .1). Further inwards from the chilled margin the very fine saccharoidal and microcrystalline character of the groundmass becomes recognisable in handspecimen (Sample 4). Microscopically the constituents of the matrix are seen to range in diameter from 0·1 to 0·5 mm. and since the average length of the phenocrysts is still under 1 cm., the symbol D.III.5 of the granite porphyry group would apply to it. As in the case of the rhyolite porphyries local coarse granite porphyry facies are also occasionally present in some of the intrusions. With further gradual increase in the grain size of the matrix the granite porphyry facies of dyke *f* grades into porphyritic granite (textural type X.III.4). Under the microscope the average grain diameter of the specimen was found to exceed 1 mm. but there is no significant increase in the average size of the insets. Dykes *e* and *m* provide good examples of intrusions consisting dominantly of granite porphyry with local gradations into coarse granite porphyries or centrally situated textural types superficially resembling granite.

D.III.6 represents the principal textural type of dykes *b*, *c*, *d*, *h*, *i*, *j*, *l*, *p* and *q*, which may grade laterally and locally into D.II.6 and occasionally also inwards into D.III.5 and D.II.5 (Fig. 6).

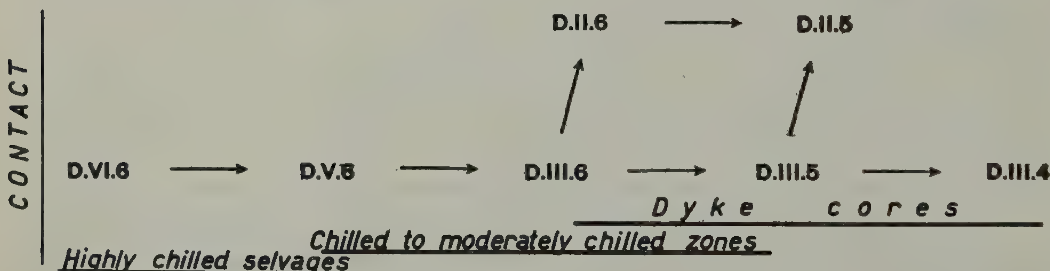

FIG. 6. Diagram illustrating the interrelation of textural types characterising the Cape St. Martin dyke rocks.

The highest degree of chilling is exhibited by dyke *a* which is incidentally also the most distant intrusion from the region of greatest dyking activity. The rock along the contact has a dense black flinty matrix (Plate III, ph. 6, sample 7) in which occasional euhedral insets about 1 cm. long and consisting of pink microcline and white plagioclase are visible. Laterally along the contact the rock may be seen to grade into the textural types D.IV.6 and D.V.6.

The chilled selvages of the intrusion *r* and dyke *k* (Plate III, ph. 7), as well as some of the apophyses, also belong to the D.VI.6 group which represents the rhyolites but is characterised by the presence of an occasional phenocryst. Microscopically the mineral constituents of the groundmass measure approximately 0·009 mm. across and small insets are sparsely distributed within it (Plate VI, pm. 6). In this rock microspherulitic to microgranophyric blebs are also present but only along the contact. The larger blebs are 0·15 mm. in diameter and they may aggregate to form a band

parallel to the contact. The band grades on either side into smaller bodies which fade out in the matrix. Some of them are round and possess a radial structure analogous to minute felspar twins or intergrowths. Under the microscope in ordinary light these bodies are indistinguishable from the groundmass, except that they appear as clear areas which either interrupt the haphazard distribution of the minute biotite tufts and scales of the groundmass, or induce the linear arrangement of them parallel to the contact. These bodies which were only observed to occur along the contact of dyke *a*, differ from the spherulites present in glassy or devitrified rocks. It would seem as if they originated after the labile shower of biotite tufts and scales, but before viscosity inhibited flowage. Their primary character considered in conjunction with the progressive increase in the crystallinity and granularity of the rocks inwards from the contact, as well as the absence of perlitic cracks, seems to indicate that the original magma which invaded the rocks in the area under consideration was never so highly chilled as to congeal in a holohyaline state. Inwards from the contact of dyke *a* the rhyolite marginal facies is seen to grade through rhyolite porphyry into granite porphyry, and this transition is accompanied by a gradual increase in the granularity of the groundmass, as in the case of dyke *f*. Rocks gradational between the two last named textural types commonly hold quartz and felspar phenocrysts surrounded by composite collars consisting of an intergrowth of felspar and quartz which may also appear as isolated bodies in the matrix of the rocks.

The rhyolites and rhyolite porphyries of the chill phase are characterised by a great variety of quartz felspar intergrowths which not only tend to grade into one another but also into the matrix of the rock. The intergrowths, which are either crypto- or microcrystalline, may be abundant, sparsely represented or absent in a given rock section. They appear in the form of bodies or patches in the groundmass or form collars about insets. Any phenocryst may act as nucleus but the collars are more frequently developed about insets of felspar. In the finest grained rocks the implication structures are plume-like or radially fibrous and may be said to become microgranophyric with slight increase in the grain size of the matrix.

In these rocks the dominant type of intergrowth is micrographic and appears to be related to the micrographic implication structure. In the immediate vicinity of phenocrysts the matrix sometimes tends to become more coarsely crystalline, while the interstitial extension of the substance of an inset may give rise to a type of microlacinear structure. Where the matrix becomes coarser the ragged edges of such phenocrysts may be seen to completely envelop optically determinable and differently orientated, rounded granules of the matrix with the production of a sieve-like border known as lacinear structure (Plate IV, pm. A3 and Plate V, pm. 2). Microcline insets may also display discontinuous myrmekitic borders.

Composite insets which may be simply twinned sometimes appear in the chill phase along the contact. These remarkable phenocrysts, which like ordinary insets may be partially or completely enveloped by the fine-grained collars described above, are unique in that they consist of a relatively coarse intergrowth which is out of harmony with the fine-grained nature of the surrounding matrix. In one of these composite crystals the strained and irregular microcline core was observed to be in optical continuity with the twinned microcline of the graphic mantle (Plate V, pm. 1). In the granite porphyry facies of the dykes the matrix is invariably characterised by the presence of patches of granophyric or graphic intergrowths, the coarseness of which is related to the grain size of the matrix. The relative proportion of normal matrix intergrowths may be said to vary from one specimen to another. By staining these rocks with sodium cobalti-nitrite, after they had been previously treated with hydro-

fluoric acid (A. Gabriel and E. P. Cox, 1929), the coarser intergrowths are made readily discernible in the matrix of a handspecimen.

In the granite porphyry of dyke *a* (Plate VII, pm. 6), as well as in certain samples from dyke *f*, the entire groundmass of the rock consists of micrographic intergrowths. As the centrally situated facies of dyke *f* is approached the entire matrix of the granite porphyry gradually becomes coarser. With the increase in grain size the graphically intergrown quartz and alkali felspar tends to lose its characteristic structure while simultaneously decreasing in amount relative to the rest of the matrix, which develops a granitoid fabric (Plate IV, pms. A5-6).

Quartz

Apart from the much cracked, rounded and irregular quartz grains up to 9 mm. in diameter, perfect euhedral and symmetrical bipyramids occasionally modified by a narrow but regular zone of prism faces, may be seen to stand out in relief on weathered surfaces of the micro- or cryptocrystalline varieties of the Cape St. Martin intrusions.

Microscopically the quartz phenocrysts (Plate VI, pm. 1) in the spherulitic matrix of the rock along the contact of dyke *a* are well rounded and show signs of magmatic corrosion. Under crossed nicols the resorbed quartz insets are seen to be enveloped by a rim of non-spherulitic groundmass which is also observed to extend into resorbtion embayments in the phenocrysts. β-quartz phenocrysts also exhibit evidence of early protoclastic deformation, since the fracture zone traversing the crystals displays evidence of conspicuous channel corrosion. Analogous phenomena have also been described by D. L. Scholtz (1946, Plates XIII and XIV, Figs. 3 and 2, and 3 respectively) in the quartz porphyries of Hoedjies Bay, Saldanha.

The preservation of the reaction product by the rapid chilling of the magma in these examples provides us with indisputable evidence of the actuality of the resorption process and the nature of the fluid medium, arrested in the act of corroding a crystalline solid. It should be pointed out, however, that some geologists believe that the embayments of quartz may be ascribed to deuteric action or the infilling of irregular magmatic cavities (G. Laemmlein, 1930, 1933). Excluding the irregular and strongly resorbed crystals β-quartz typically appears in the form of perfect idiomorphic phenocrysts in the rhyolite facies of the intrusions, but in the coarser textural types the insets gradually lose their identity by extensive re-crystallisation and cannot be distinguished from interstitial quartz of late crystallisation.

Anhedral mineral and liquid inclusions in quartz are distributed at random or arranged in sheets. The larger inclusions comprise minerals such as biotite, muscovite, apatite, allanite, fluorite, zircon and a few others which could not be positively identified. Very prominent in the quartz crystals are the sheets of fluid inclusions arranged in parallel sets which show a high degree of preferred orientation in thin section. Some of these sheets of liquid inclusions are sinuous or curved and others display en echelon or partly radial distribution. Bubbles showing Brownian movement are present in many of the inclusions.

The crystallographic orientation of the planes of fluid inclusions in quartz was determined on the Universal stage in the conventional manner. In single thin sections the poles normal to the planes of liquid inclusions tended to cluster, but the c-axes of the quartz phenocrysts showed no preferred orientation in the diagram. Petrofabric investigations of a dyke contact revealed the presence of one and two sets of sheet inclusions in the intrusive and granite wall rock respectively. The maximum of the former set, which was orientated normal to the plane of the contact, was found to coincide with one of the two set maxima of the wall rock. It was found that the

planes of inclusions are in no way related to the structure of the quartz crystals. This is also borne out by the fact that parallel sets of inclusions are seen to cross haphazardly orientated β-quartz crystals in a single section (Plate VI, pm. 2).

O. F. Tuttle (1949) made an extensive study of planes of fluid inclusions and distinguished primary and secondary groups. He thinks that secondary planes of fluid inclusions "are believed to have originated as fractures in quartz grains, which subsequently became filled with solution. The solution-filled fracture is believed to change to numerous individual inclusions by a process of solution and deposition of silica. The process may have originated by a mechanism similar to that discussed by Riecke (1895), which requires that solution take place in a stressed crystal in the area under greatest stress and precipitate in the area of least stress".

T. N. Dale (1923) quoted by Tuttle, believes that sheets of inclusions in granite are largely primary in origin.

A study of the protoclastic features exhibited by the dyke rocks proved to be of considerable interest in this problem. In the chilled margin of dyke *f* the rare phenomenon of "pre-consolidation microshearing" was recognised (Plate VI, pms. 4 and 5). By examining the continuation of the biotite-lined proto-microshear under crossed nicols, one finds that it does not interrupt the interlocking texture of the granular components of the matrix. A quartz phenocryst lying across its path shows a minute offset and beyond this crystal the protoshear is seen to terminate against the next inset of β-quartz, which is traversed by numerous parallel planes of fluid inclusions. Some phenocrysts show signs of selective resorption along such directions of microshearing, which indicates that they must have originated before complete consolidation of the matrix. The presence of fluorite is occasionally observed along them. These protoshears are normally very short and impersistent, rarely attaining lengths of 2 cms., and this type of deformation appears to have produced the columnar variety of undulose extinction observed in some phenocrysts. The earlier and shortest are of variable orientation, or even curved, whereas the longer and later protoshears are parallel to the preferred orientation of planes of fluid inclusions in a given rock slice. F. G. H. Blyth (1949) described a similar type of shearing which did not affect the groundmass of porphyrites. He therefore concluded that these originated during the consolidation of the rock.

The fact that the outer extremity of a fracture in a quartz phenocryst has been selectively subjected to channel resorption while the inner extremity peters out and is replaced by a plane of fluid inclusions which in turn disappears within the body of the crystal, forces one to conclude that the planar distribution of the fluid inclusions constitutes a primary feature of the rock. This *protoshear hypothesis* for the origin of planes of liquid inclusions is also corroborated by other observations. A quartz phenocryst examined under crossed nicols displayed a wedge-like sector slightly out of optical continuity with the rest of the crystal, but in ordinary light the form of the wedge was outlined by planes of liquid inclusions restricted to the substance of the crystal. The early origin of these fluid inclusions is also verified by the fact that cavity resorption of the phenocrysts is observed where closely spaced planes of such inclusions intersect margins of insets.

Because quartz is a very "sensitive mineral to deformation" (F. J. Turner, 1948), it becomes more and more re-crystallised (Plate VI, pm. 3) towards the core of the dyke and is eventually disguised by late interstitial quartz and insertal fabric, so that it can no longer be recognised as the original β-quartz except for the occasional inclusions of muscovite. Occasionally, re-crystallised phenocrysts may even be present in the chilled margins of the dyke (Plate VI, pm. 2). They tend to retain their primary

planes of fluid inclusions which cannot be distinguished microscopically from possible post-consolidation (secondary) fluid inclusions, since they transgress sutured boundaries of the interlocking grains of re-crystallised quartz. Some planes of inclusions become disrupted, with the result that the inclusions become more scattered during re-crystallisation of the host. Irregular and angular fragments of clear, unstrained quartz devoid of fluid inclusions no doubt represent the products of the early protoclastic deformation of suspended phenocrysts in a medium whose viscosity was still sufficiently low to permit it to flow around and completely envelop them.

During the pneumatolytic stage, fractures in the late chilled veinlets present in the dykes *f* and *h* were infilled by fluorite, with or without scales of biotite. Where these fractures traverse quartz phenocrysts the mineral is seen to be locally re-crystallised, with the production of a fine mosaic ribbon of granular quartz.

Plagioclase

Plagioclase crystals range in length from a fraction of a millimetre to 3 centimetres, but the phenocrysts of 3 centimetres are rarely encountered. The average size of the large plagioclase phenocrysts is about two centimetres and these are more numerous in rocks of the southern group of dykes. The vast majority of plagioclase crystals in these intrusions do not exceed 3 mm. $\times$ 5 mm. Plagioclase, which is usually less abundant than microcline, constitutes 20 per cent by volume of the granitic portion of dyke *f* but in the chilled margin of dyke *a*, of which the groundmass comprises 94 per cent by volume, only 4 per cent of this mineral is present.

The large phenocrysts possess complex internal structures and may have broad $Ab_{75}An_{25}$ outer zones which envelop twinned and saussuritised cores holding from 35 to 45 per cent An and showing frequent resorbtion and repair. Occasional plagioclase with well defined outlines resembling those of a single crystal were found to consist of an aggregate of smaller crystals, each characterised by its own zonal pattern: some of these crystals are optically discontinuous with the rest.

By utilising composition planes, cleavages and crystal faces, the crystallographic orientation of the optical indicatrix was determined by means of the U.M. stage. Most of the values obtained in this manner corresponded with the relevant curves of Nikitin, while the anorthite content thus deduced was found to correspond closely to the composition interpolated from the axial angles which were obtained by the direct measurement of 2V in sections exhibiting both optic axes on the Universal stage. In dykes *f* and *a* the average ranges in composition of the plagioclase were found to be from An_{16} to An_{24}, and from An_{25} to An_{35} respectively.

No $2V\alpha$ value lower than 81 degrees was found in the plagioclases with an anorthite content ranging from 10 to 20 per cent. These An values were checked according to the method of R. W. Foster (1955), by measuring the refractive indices of the plagioclase glasses prepared from some phenocrysts in the chilled selvages of the dykes. From the M. L. Keith and O. F. Tuttle (1954) diagram the felspars mentioned above appear to be the normal low temperature variety of plagioclase. Attention should, however, be drawn to certain irregularities which were apparent, particularly in some of the more basic plagioclases, even though the observed anomalies may lie within the range of experimental error. In the cores of zonal crystals in which the anorthite content exceeds 35 per cent it was found that the (010) poles of different parts of the untwinned crystals or the components of twinned crystals, tended to migrate perpendicularly across Nikitin's $\perp$(010) curve towards the high temperature plagioclase curve of G. van der Kaaden (1951), whereas no such deviation characterised the more albitic mantles. These anomalies therefore, appear to be analogous to those already described by F. J. Turner (1947) and O. Bradley (1953). In

other phenocrysts brachypinacoidal poles of the cores were observed to occupy the region between the above-mentioned Nikitin and van der Kaaden curves. A third type of discrepancy was observed to occur between the anorthite content deduced from Nikitin's axial angle and $\perp$(010) curves for plagioclase having the composition Ab_7An_3, if the latter be assumed to be the more accurate of the two. In the Keith and Tuttle (1954) diagram this type of felspar is located in the field immediately below the junction of the high and low plagioclase curves.

The optical characteristics of the many plagioclase crystals that were studied seem to indicate that they are of the low temperature variety. The cores of zonal crystals with an anorthite content exceeding 30 per cent may represent transitional types (O. Bradley, 1953) between high and low temperature plagioclases.

The plagioclase, in order of abundance, is twinned according to the Albite, Carlsbad, Albite Ala and Manebach laws, but Roc Tourné and Baveno twins were also found. In the chilled selvages polysynthetically twinned crystals show no interruption of the zoning, and this indicates that the twinning is syngenetic. Narrow but well spaced twin lamellae of late origin (R. C. Emmons and V. Mann, 1953), were, however, also observed.

The zoning in plagioclase was also found to be normal continuous or discontinuous, and reversed. Highly zoned crystals, especially of the rhythmic type, are most abundant in contaminated rocks of the southern group of dykes. In these intrusions, and more particularly in dyke *a*, the highly chilled margin contains small phenocrysts (1·5 mm. in length) with relatively basic cores ranging in composition from An_{40} to An_{60}. Their clear and sharply defined euhedral cores, which exhibit normal continuous zoning, may easily be recognised by the displacement of the Becke line owing to the lower 20 to 25 per cent anorthite content of the narrow mantles. Such basic plagioclase crystals were not observed in the rocks of dyke *f* or in the form of inclusions in phenocrysts of microcline.

Sericitisation and saussuritisation of the cores are typical of the larger phenocrysts. In the chilled selvage of dyke *a* almost completely sericitised plagioclase is found together with unaltered phenocrysts. Such alterations as well as the clouding, though clearly related to the zoning of the felspar, was also observed to transgress zones of cored crystals, and is, therefore, not always primarily controlled by the chemical composition.

In the chilled selvages the smaller phenocrysts which have been partly sericitised and often possess irregular cracked cores, are seen to be rimmed by clearer outer mantles of the same composition (Plate VII, pm. 1). Apart from the possibility of selective sericitisation, this structure and the intimate association of sericitised and unaltered phenocrysts in the chill phase suggests an earlier epoch of deuteric alteration. Many of the undoubted plagioclases are pinkish in colour and can only be distinguished in handspecimen from the microcline by means of the staining process previously described.

Microcline-microperthite

The pink to brick-red euhedral insets of microcline-microperthite are conspicuous in the bluish grey matrix of the rhyolite porphyry facies of the Cape St. Martin dykes and contribute in no small measure to the handsome appearance of the rock. Two generations of phenocrysts, whose average dimensions are 3 cms. × 1·8 cms. and 0·8 cms. × 0·6 cms., are present in all the textural rock types. Like plagioclase both large and small insets not only increase in abundance, but also slightly in size as the core of an intrusion is approached. This fact, considered together with the

presence of large microclines in the narrow highly chilled selvages, is clearly indicative of the pre-intrusion crystallisation of the dyke magma.

The phenocrysts which display faces of the crystal forms $\{010\}$, $\{1\bar{3}0\}$, $\{1\bar{1}0\}$, $\{\bar{1}11\}$ and $\{\bar{2}01\}$, may be equant, elongated in the direction of the c-crystallographic axis, or more rarely tabular parallel to (010). Though megascopic inclusions of biotite and grains of white plagioclase are commonly visible in phenocrysts of microcline, the reverse relation was rarely observed to hold in the two last named minerals.

Microscopically the potash felspar which very rarely exhibits gridiron twinning is also seen to be perthitic. Simple Carlsbad twins are common in all varieties of microcline, but twinning according to the Baveno law is relatively uncommon. In the larger and generally somewhat cloudy phenocrysts the fine perthitic segregations parallel to $(150\bar{2})$ are characteristic, while in the clear interstitial microcline of later crystallisation only well defined stringers and rods of oligoclase are to be seen. Phenocrysts may occasionally be zoned or may exhibit relic structures suggestive of zoning. Insets characterised by an optically homogeneous potash felspar base or host, may exhibit localised strained areas, while other crystals show undulose extinction which may partially obliterate the zonal structures of cored crystals.

Zoned phenocrysts and microcline displaying cross-hatching occur in dykes *a* and *i* respectively. The other varieties of alkali felspar referred to are typical of the differentiated intrusions in which interstitial microcline increases in amount with the increasing granularity of the rock.

The negative optic axial angle of microcline was found to vary from 88 degrees to 70 degrees with peaks at the 2V values of 83, 76 and 74 degrees. The size of the axial angle of microcline does not appear to be related in any way to the distance of the mineral from the contact, but in the largest, and therefore, the earliest phenocrysts $2V\alpha$ is usually 76 to 80 degrees.

Analogous variation in the size of the optic axial angles of the microcline-microperthite from the George granite batholith were attributed to strain effects by C. T. Potgieter (1950, p. 372). L. Dolar-Mantuani (1952) also found a wide range of the 2V values returned by microcline in the Westkettle rocks. No regular relation between the size of the axial angle and the amount or type of exsolved plagioclase in the case of clear microcline host crystals was found to exist, and it is felt that even if this line of investigation had yielded positive results, the reliability of the determinations would still have been open to doubt unless the proportion of submicroscopic segregations had been taken into account.

Subradial exsolution of perthitic material around inclusions of quartz can only be attributed to local strain in the host caused by the inclusion. This observation suggests that stress as well as temperature should be taken into consideration in accounting for the origin of exsolved perthites. In general the effect of strain on phenocrysts may be recognised by the wavy habit of the $(150\bar{2})$ exsolution lamellae. From a study of perthites associated with undulant quartz in micaceous calc-alkaline granites, F. Chayes (1952) concluded that perthite "has been formed by exsolution from an originally (optically) homogeneous microcline, and that its formation is favoured, if not actually induced, by shearing stress".

The effect of strain on the crystallographic orientation in microcline is unfortunately not known. Measurements conducted on partially strained phenocrysts on the U.M. stage seem to indicate a reduction in triclinicity of otherwise normal microcline in the strained areas. It is, therefore, impossible to ascertain to

what extent the undulatory extinction of some phenocrysts is governed by difference in composition or the effects of strain.

There is little or no difference between the nature and abundance of the exsolved plagioclase in the microcline from the intrusions and the wall rock. In some dyke microclines, the perthitic material may be abundant while other phenocrysts are almost clear and devoid of microscopically visible segregations of plagioclase. The crystallographic orientation and different types of regular perthitic material found in microcline may be tabulated as follows in decreasing order of abundance (Plate VIII, pms. 1 and 2):

$(150\bar{2})$	stringlets ($\pm$ 0·008 mm. and less).
(varies)	irregular strings simulating vein perthite ($\pm$ 0·02 mm. in width).
(001)	strings and stringlets.
(010)	strings and rods.
$(1\bar{3}0)$	strings and rods.
$(1\bar{1}0)$	strings and rods.

Irregularly distributed bead-like segregations, which avoid the rims of phenocrysts, were also observed. The strings are short in comparison to the $(150\bar{2})$ stringlets, which often traverse the full width of the host and the latter is characteristic of the early large phenocrysts. The stringlets become constricted before reaching the rims of a microcline phenocryst but wedge out in the vicinity of irregular perthitic material.

Fig. 7. Variations in the optical properties of microcline exhibited by two sections of a zoned phenocryst, twinned according to the Carlsbad A law.

Zonal crystals of microcline-microperthite were observed in dyke *a*. That such zonal phenocrysts are also present in the highly chilled margin of this intrusion strongly suggests that they were present in the magma at the time of emplacement. The zonal structure is best developed in the outer half of any phenocryst and individual zones are usually narrow and sharply defined. The zoning is normally discontinuous or even reversed in relation to triclinicity. In zonal crystals displaying local undulose extinction, differences in the optical behaviour of zones in the strained portions of the crystals are levelled out (Plate VIII, pm. 3).

In certain microcline-microperthite crystals the selective concentration of kaolinised exsolved plagioclase in alternate zones, results in the production of pseudo-zoning. This phenomenon is well displayed in the upper half of the twin in the photomicrograph (Plate VIII, pm. 5), whereas the lower half shows complete extinction even though characterised by a similar structure.

Figure 7 illustrates two different sections of the same phenocryst taken ten inches from the contact of dyke *a*. This crystal contains numerous exsolved ($150\bar{2}$) stringlets of plagioclase some of which are seen to peter out before reaching the outermost narrow zone. Irregularly distributed perthitic material concentrated in the vicinity of small inclusions of plagioclase and quartz were observed to intersect the internal zonal structure as well as the mantle boundary of the host crystal. In the diagram (Fig. 7) it will be seen that the extinction and axial angles of different parts of the narrow mantle vary considerably, but in the inner zones which are not so continuous or well defined, the variation is not so great. Measurements conducted on the Universal stage show that $\gamma/\perp(010)$ ranges from 5 to 18 degrees, which is beyond the possible experimental error.

In another microcline-microperthite phenocryst the first and third layers from the centre of this zonal crystal returned similar values, namely, $2V\alpha = 76°$, $\gamma/\perp(010) = 9°$, whereas the second or middle zone was characterised by the following constants: $2V\alpha = 83°$, $\gamma/\perp(010) = 18°$.

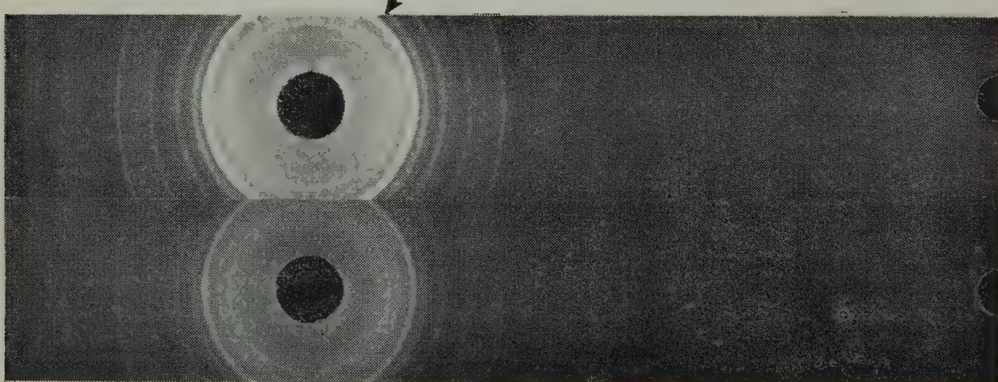

FIG. 8. X-ray powder diffraction photographs, Cu Kα radiation showing diffused nature of (131) lines of: a. Zonal microcline phenocryst illustrated in Fig. 7. b. Unzoned microcline also from dyke a.

In short, the optical properties of zonal microcline-microperthite suggest a distinct variation in the degree of triclinicity of the host crystal or its component parts, and this variation is corroborated by the fact that X-ray diffraction photographs (Fig. 8) of powders taken from different zonal and unzoned crystals all display a single broad diffuse and irresolvable (131) reflection (J. R. Goldsmith and F. Laves, 1954b).

Biotite

Brown euhedral crystals of biotite, which measure two millimetres in diameter, are present in all rock types. The biotite, partly or completely altered to chlorite is usually associated with epidote. Local nests of biotite are intimately associated with small rounded grains of quartz and accessory minerals of earlier crystallisation. These quartz granules, which hold numerous apatite needles, may also occasionally include zircon and rounded grains of tourmaline. The fact that the inclusions of the accessory minerals are concentrically arranged in six-sided biotite plates, indicates that they were probably mechanically enveloped during crystal growth. Magnetite crystals appear within the larger biotite phenocrysts, which are usually dark brown to reddish brown in colour.

Biotite tufts and scales in the chilled margins of the dykes tend to form clusters (Plate VI, pm. 5) and conspicuous collars about flakes and biotite phenocrysts. It is possible that these scales and tufts of biotite represent labile products of crystallisation owing to the supercooling of the dyke magma at the time of intrusion. The fact that they increase in size from the margin towards the centre of dyke *f* (Plate IV, pms. B) seems to indicate that they constitute the nuclei of a second generation of biotite.

In the granophyric core of dyke *a* the biotite forms straight or crooked or film-like aggregates which may partially envelop or separate crystals of quartz and felspar (Plate V, pm. 4). The fabric of the micrographic intergrowths of quartz and alkali felspar may also be modified by the presence of these film-like aggregates and scales of biotite, which appear to have been concentrated as a result of the marginal or radial expulsion of pre-existing tufts and scales of biotite, during the formation of the intergrowths by the force of crystallisation of quartz and felspar.

Muscovite

Tiny flakes of white mica presumably muscovite with $\gamma - \alpha = 0{\cdot}034$ have been found to occur only in β-quartz phenocrysts. D. L. Scholtz (1946, p. lix) in referring to the Cape Granites, states: "Very little if any of the muscovite appears to be of primary origin". According to J. P. Iddings (1909, Vol. I, p. 131) muscovite may be of pyrogenetic origin in rocks crystallising under considerable pressure, while F. W. Clarke (1920) believes that primary muscovite is essentially a mineral constituent of deep-seated rocks like granite and quartz porphyries. The fact that fluorite is often present about the margins of the muscovite flakes when considered together with the possibility that fluorine ions may proxy for the (OH)-radical in micas, may be related to the abnormally early crystallisation of muscovite in the rocks under consideration.

Accessory minerals

A grey to reddish brown isotropic mineral substance was found within crystals of biotite or surrounded by tuft-like aggregates of that mineral, in the matrix of the chill selvages of the Cape St. Martin dykes. Unaltered remnants of an anisotropic and pleochroic mineral proved the isotropic substance to be pseudomorphous after allanite with $2V\alpha = 46°$, $\gamma/c = 32°$, and α = pale yellow, β = light orange, γ = yellowish brown. Rusty brown stains were observed to follow cracks and discolour the adjacent crystals and rock matrix in the immediate vicinity of the secondary

substance as a result of further decomposition. It is noteworthy that, contrary to the views held by some metasomatists, allanite is not only a magmatic product but also a product of early crystallisation.

Up to 0·3 per cent by volume of colourless fluorite occurs in the chilled phase of dyke *f*, either as an interstitial component of the groundmass or in the form of isolated crystals surrounded by relatively coarser grained matrix. Fluorite was also observed to be associated with the protoshears previously described.

Secondary calcite, when present, is restricted to the saussuritised cores of plagioclase.

Euhedral zircons terminated by pyramidal faces are found in biotite and rarely in quartz.

Slender needles of apatite occur in biotite, β-quartz and felspars, usually being orientated parallel to the crystal faces of the phenocrysts. Apatite crystals were also observed in the groundmass of the chilled selvages of the intrusions.

B. Chemistry of the Dykes

The differentiation and course of crystallisation of the intrusive dyke magma at Cape St. Martin can be conveniently discussed by focussing attention on dyke *f*, which incidentally also displays the full sequence of texturally different varieties of dyke rocks obtaining in this area.

The samples 1 to 5 were taken from the margin inwards and sample 6 was taken from the country rock adjacent to this dyke. Dyke *a* was selected because of its contamination by the hybrid granite and its having a chilled selvage showing the highest degree of marginal chilling in the area. Specimens 7 and 8 represent respectively the chilled margin and the granophyric core of dyke *a*. Four more analyses have been included in Table I*a*: the hybrid microgranite, two miles south of the area mapped; the quartz porphyry from Saldanha Bay; the younger medium even-grained granite, Cape Castle; the older coarsely porphyritic granite comprising the Saldanha batholith of this area.

The specimens for analysis were taken on fresh rock outcrops away from crush and shear zones in dyke *f* where the effects of chilling do not fade out too rapidly. The samples 1 to 5 for chemical analysis were also selected on grounds of a preliminary modal determination to justify the chemical analysis of the different textural types. Numbers 1, 2 and 5 are analyses of single specimens, while numbers 3 and 4 are composite samples taken over broader zones of 20 inches and 24 inches respectively. The modes (Table II) were selected in the same way, with numbers 3 and 4 representing composite modes of individual specimens.

Owing to the porphyritic nature of these rocks the writer first of all attempted to integrate square foot blocks on a megascopical scale in the field. The measurement of such uneven surfaces, however, proved unsatisfactory and the scheme was dropped and the mode was determined in the following manner. The average grain size of phenocrystic material is larger than 2 mm. and there was no need to stain these samples, as the flesh coloured microcline could be easily distinguished from the white plagioclase. On a polished handspecimen with a laid out grid interval of 3 mm. over an area of 200 to 300 square millimeters the components quartz, plagioclase, microcline and biotite larger than 1 mm., were readily determinable megascopically. The rest being classified as matrix. In order to correct the analyses with regard to the microscopically decipherable constituents, this matrix was subjected to a micrometrical modal analysis with the aid of an integrating stage, making north-south and east-west traverses, spaced 1 mm. apart. This served to resolve the megascopic ground-

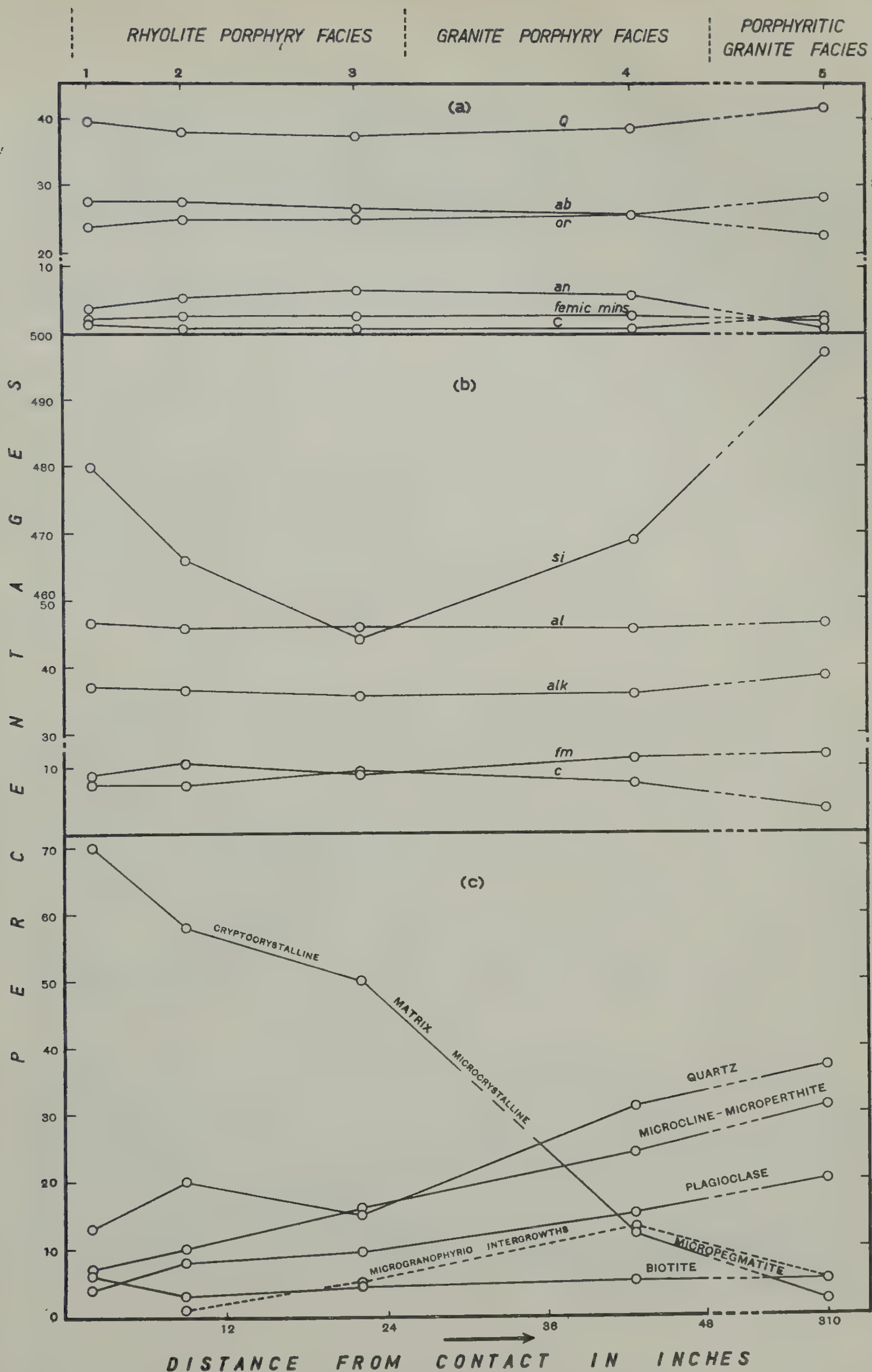

FIG. 9. Differentiation diagrams of dyke *f*.
(a) Normative minerals.
(b) Niggli values.
(c) Modal minerals.

mass into microscopically recognisable mineral grains and undecipherable matrix. By combining the megascopical and micrometrical results, the final mode of the rock was arrived at. To make this calculation, two micrometrical modes of the same rock slide were necessary, in one of which only the volumetric proportions of the constituents larger than 1 mm. were determined (similar to the megascopical mode). The measurements were conducted in a single plane as the specimens displayed no preferred orientation.

In the chilled part of dyke *f* a separate determination on the minute irregular biotite tufts (Plate IV, pm. B1-2) was made and added to the weight percentage of the biotite flakes and phenocrysts at the expense of the weight percentage of the groundmass.

A glance at Table I*a* reveals that the dyke rocks are closely similar in composition, and various diagrams have been utilised to illustrate the slight variations in the chemical and mineralogical composition of these rocks. If the Niggli values plotted against the distances from wall to the centre of dyke *f* are used, the *si* figures immediately draw attention because they have the highest values in the margin and core (Fig. 9b). The curves *alk* and *al* are mutually sympathetic; *alk*, however, shows a slight enrichment towards the core of the dyke. The *fm* and *c* curves have antipathetic relations showing a notable increase in *fm* and a decrease of *c* towards the granitic part of the intrusion. The value *k* (Table I*b*), which represents the ratio of the potash to total alkalis, shows a slight but perceptible increase up to 3 feet 6 inches from the contact (No. 4) and is afterwards followed by a slight decrease to the core of the intrusion.

The volumetric modes plotted against the distance from the margin to the core of the dyke show a progressive increase in the quantity of all essential mineral constituents (Fig. 9c). It will be observed that the curves for plagioclase and microcline-microperthite are sympathetic, while the proportion of quartz also show a general increase analogous to the felspar components of the rock. Within the first nine inches from the dyke margin, however, free quartz shows a local increase of 7 per cent by volume (Fig. 9, mode 2), which is offset by a corresponding local decrease of the groundmass value if the modes of the other minerals are taken into account. From mode number 2, the volumetric percentage of the microgranophyric intergrowths is observed to vary antipathetically with the matrix to a maximum of 13 per cent in number 4. About three feet from the contact the matrix, in samples 4 and 5, becomes coarse enough to be microscopically determinable (Plate IV, pms. A5). For convenience it has been graphically represented in Fig. 9b as micropegmatite which decreases in quantity towards the core of the intrusion. The proportion of the biotite, on the other hand, is subjected to little variation, apart from a slight increase in volumetric percentage in the immediate vicinity of the contact.

When the curves representing the Niggli values and the modes are compared, the high *si* value of the chilled margin is revealed by the highly siliceous character of the abundant matrix holding 13 per cent by volume of β-quartz phenocrysts, while in the central and granitic portion of the dyke it is mainly due to the presence of the recrystallised and interstitial quartz (Plate IV, pm. A6). The lowest *si* value corresponds to the depression of the modal analysis curve representing phenocrystic quartz. This depression in the modal curve can be explained by a slight decrease in the silica content by differentiation at this stage, as well as a notable increase in the volumetric percentage of the microgranophyric intergrowths. The peculiar trend of this modal curve may indicate a stage of undercooling.

TABLE Ia

No.	SiO_2	Al_2O_3	Fe_2O_3	FeO	MgO	CaO	Na_2O	K_2O	H_2O	H_2O	TiO_2	P_2O_5	MnO	CO_2	S	F	Cl	Total
1	76·10	12·47	0·88	0·57	0·12	1·08	3·40	4·00	0·32	0·04	0·01	0·00	0·04	0·21	0·05	0·08	0·15	99·52
2	76·31	12·84	0·88	0·43	0·37	1·12	3·37	4·20	0·37	0·09	0·01	0·00	0·03	0·00	0·04	0·07	0·16	100·29
3	75·24	13·01	0·80	0·57	0·29	1·44	3·29	4·20	0·36	0·06	0·01	0·00	0·04	0·06	0·03	0·08	0·16	99·64
4	76·00	12·51	0·96	0·57	0·41	1·12	3·15	4·28	0·27	0·03	0·01	0·00	0·02	0·02	0·04	0·01	0·15	99·55
5	76·63	12·08	1·28	0·43	0·30	0·52	3·64	3·88	0·47	0·07	0·01	0·00	0·03	0·27	0·04	0·04	0·27	99·96
6	76·33	12·71	0·66	0·43	0·33	1·06	3·51	4·00	0·40	0·06	0·01	0·00	0·02	0·00	0·08	0·04	0·15	99·79
7	76·40	13·13	0·72	0·93	0·22	1·12	3·37	4·02	0·52	0·04	0·02	0·01	0·04	0·00	0·00	0·06	0·25	100·85
8	74·86	13·22	0·96	1·29	0·12	1·24	3·15	3·93	0·56	0·08	0·03	0·04	0·05	0·00	0·02	0·08	0·16	99·79
9	70·26	14·50	0·64	2·01	1·20	2·80	3·47	3·53	0·76	0·11	0·50	0·09	0·05	n.d.	n.d.	n.d.	n.d.	99·92
10	74·82	12·36	0·00	2·50	0·10	1·47	3·20	5·01	0·29	0·05	0·21	0·17	n.d.	0·04	n.d.	n.d.	n.d.	100·22
11	76·22	12·39	0·80	0·72	0·43	1·01	3·67	4·67	0·26	0·11	0·13	tr	tr	n.d.	n.d.	n.d.	n.d.	100·41
12	71·40	13·41	0·62	2·67	0·45	2·58	3·62	4·41	0·41	0·09	0·40	0·19	0·04	0·06	n.d.	n.d.	n.d.	100·07

TABLE Ib

No.	*si*	*al*	*fm*	*c*	*alk*	*mg*	*c/fm*	*k*	*al-alk*	*qz*	*h*	*co₂*	*s*	*fr*	*cl*	Q	L	M
1	480	46·6	8·7	7·6	37·1	0·13	0·87	0·44	3·9	231	6·4	1·9	0·7	1·5	1·5	60·6	36·9	2·5
2	466	45·8	10·3	7·3	36·6	0·32	0·71	0·45	4·0	219	7·3		0·4	1·1	1·5	59·9	37·3	2·8
3	454	46·0	9·1	9·4	35·5	0·28	1·04	0·46	3·4	211	8·0	0·4	0·4	1·4	1·4	59·7	37·9	2·4
4	469	45·6	11·1	7·4	35·9	0·33	0·67	0·47	3·7	224	5·2	—	0·4		1·4	60·4	36·5	3·1
5	497	46·3	11·7	3·5	38·5	0·30	0·30	0·41	4·9	241	10·1	2·3	0·4	0·8	3·1	61·0	35·4	3·6
6	480	47·2	8·3	7·2	37·3	0·36	0·86	0·43	3·8	229	8·3	—	0·7	0·7	1·5	60·6	36·9	2·5
7	465	46·7	10·2	7·3	35·8	0·21	0·72	0·44	4·3	220	10·6	—	—	1·1	2·6	60·2	36·5	3·3
8	452	46·8	12·3	7·6	33·3	0·09	0·62	0·45	2·5	216	11·2	—		1·4	1·4	60·5	35·3	4·2
9	333	40·3	19·0	14·2	26·4	0·45	0·75	0·40	1·9	127	11·6	—	—	—	—	53·9	39·6	6·5
10	430	42·1	12·7	9·0	36·2	0·05	0·72	0·50	6·1	185	5·5	0·3				57·4	38·4	4·2
11	454	43·6	11·1	6·4	38·9	0·35	0·58	0·46	8·3	198	5·0					59·1	38·2	2·7
12	352	38·7	16·6	13·6	31·1	0·20	0·82	0·45	4·0	128	6·5	0·3	—	—	—	53·2	40·1	6·7

TABLE Ic

No.	Q	or	ab	an	C	h FeO.SiO₂	h MgO.SiO₂	mt	ht	cc	hl	fr	Total	C.I.P.W. Symbol	Shand Symbol
1	39·66	23·91	27·77	3·61	1·43	0·30	0·26	1·39	—	0·50	0·23	0·16	99·22	I 323	DOpa6
2	37·92	25·02	27·77	5·28	0·82	0·90	0·13	1·39	—	—	0·23	0·07	99·53	I 323	DOpa3
3	37·32	25·02	26·72	6·39	0·82	0·70	0·40	1·16	—	0·10	0·23	0·16	99·02	I 323	DOpa5
4	38·64	25·58	25·68	5·56	0·82	1·00	0·26	1·39	—	—	0·23	—	99·16	I 323	DOpa5
5	41·76	22·80	28·30	0·56	2·24	0·70	—	1·62	0·16	0·60	0·47	0·07	99·29	I 313	XOpa5
6	38·64	23·91	28·30	5·00	1·02	0·80	0·26	0·93	—	—	0·23	0·07	99·16	I 323	XOp (βL)
7	39·00	23·91	27·25	5·28	1·43	1·50	1·32	0·93	—	—	0·35	0·16	100·13	I 323	DOpa1
8	39·30	22·80	25·68	5·28	2·04	1·30	1·58	1·39	—	—	0·23	0·16	98·76	I 323	DOp(βL)

TABLE II

No.	Quartz	Microcline	Plagioclase	Biotite	Groundmass	Micrographophyric intergrowths and micropegmatite	Total	Textural Symbol	Shand Symbol
1	13	7	4	6	70	—	100	D.III.6	DOpa6
2	20	10	8	3	58	1	100	D.III.6	DOpa3
3	15	16	9·5	4·5	50	5	100	D.III.6	DOpa5
4	31	24	15	5		25	100	D.III.5	DOpa5
5	37	31	20	5		7	100	X.III.4	XOpa5
7	1	0·5	4	1	94		100	D.VI.6	DOpa1

Analysis No.	Rock Type		Locality	Analyst
1*	Rhyolite Porphyry	D.III.6	Northern chilled margin, 125 feet due west of southern beacon, dyke *f*, Cape St. Martin.	P. R. B. Heyman.
2*	Rhyolite Porphyry	D.III.6	Sample taken 9 inches from the contact south of No. 1, dyke *f*, Cape St. Martin	P. R. B. Heyman
3*	Rhyolite Porphyry	D.III.6	Composite sample 20 inches, taken 19 inches from the contact, south of No. 1, dyke *f*, Cape St. Martin.	P. R. B. Heyman
4*	Granite Porphyry	D.III.5	Composite sample taken 3 feet 6 inches from the contact, south of No. 1, dyke *f*, Cape St. Martin.	P. R. B. Heyman
5*	Porphyritic medium even-grained Granite	X.III.4	Granitic core of dyke *f*, Cape St. Martin	P. R. B. Heyman
6*	Medium even-grained Granite	X.VI.4	Sample 2 feet from the northern contact, dyke *f*, Cape St. Martin.	P. R. B. Heyman
7*	Rhyolite	D.VI.6	Selected sample with no visible phenocrysts, from feebly porphyritic chilled margin of the north-western contact, dyke *a*, Cape St. Martin.	P. R. B. Heyman
8*	Micro-granophyre	D.III.5	Sample from ± 15 feet granophyric core, dyke *a*, Cape St. Martin.	P. R. B. Heyman
9	Hybrid microgranite	D.VI.5	Great Paternoster	C. J. Liebenberg
10	Quartz Porphyry	D.III.6	Hoedjies Bay Peninsula, Saldanha	D. L. Scholtz.
11	Medium even-grained Granite	X.III.5	South of Cape Castle	C. J. Liebenberg.
12	Coarsely porphyritic Granite	X.II.4	Vredenburg	D. L. Scholtz

* New analyses.

The progressive increase in the amount of alkali felspar and plagioclase and sympathetic relations of their modal curves indicate that the relative proportions of these two felspars remain more or less constant, while the later antipathetic trend of *alk* and *c* discloses a simultaneous tendency to the enrichment of the plagioclase in the albite component from the margin to the core of the intrusion. The fact that the quantity of albite exceeds that of orthoclase in the norms (Fig. 9a), whereas microcline-microperthite is more abundant than the plagioclase in the modes of these rocks, clearly points to the sodic nature of the microcline apart from the exsolved plagioclase.

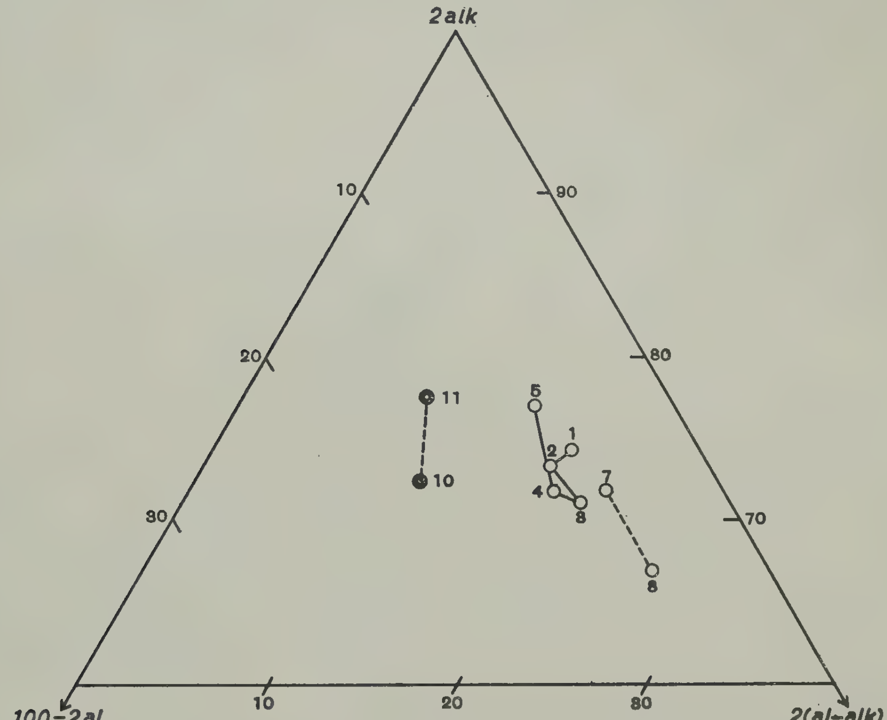

FIG. 10.

The nature of the differentiation process in dyke *f* may be illustrated by means of the ternary diagram (Fig. 10) in which the Niggli values 2*alk*, 2(*al-alk*) and 100-2*al* approximately reflect the relative proportions of the normative felspars and dark minerals in the analysed samples 1 to 5 (Table I*a* and I*b*). Inwards from the margin of the dyke, that is, across the highly chilled portion of the intrusion, a slight increase in the normative anorthite content is perceptible. Further inwards, with increase in crystallinity and granularity towards the centre of the intrusion (Analyses 3 to 5), a notable increase in the amount of alkali felspar is seen to occur at the expense of anorthite.

The Q.L.M. diagram (Fig. 11) founded on the basic molecular norms, indicates no variation in the composition of the felspars, since the normative quartz content of the rock is taken into account. In this ternary diagram it will be seen that from

samples 1 to 3 across the outer chilled portion of the intrusion the quantity of normative felspar as a whole increases slightly, primarily at the expense of quartz, after which the process is reversed with very slight increase in the proportion of mafites.

In short, from the chemical aspect the differentiation of the dyke magma on consolidation involves a decrease in quartz together with an increase in the amount and lime content of the felspar, followed by an increase in the quartz content primarily at the expense of felspar of an increasingly alkaline character.

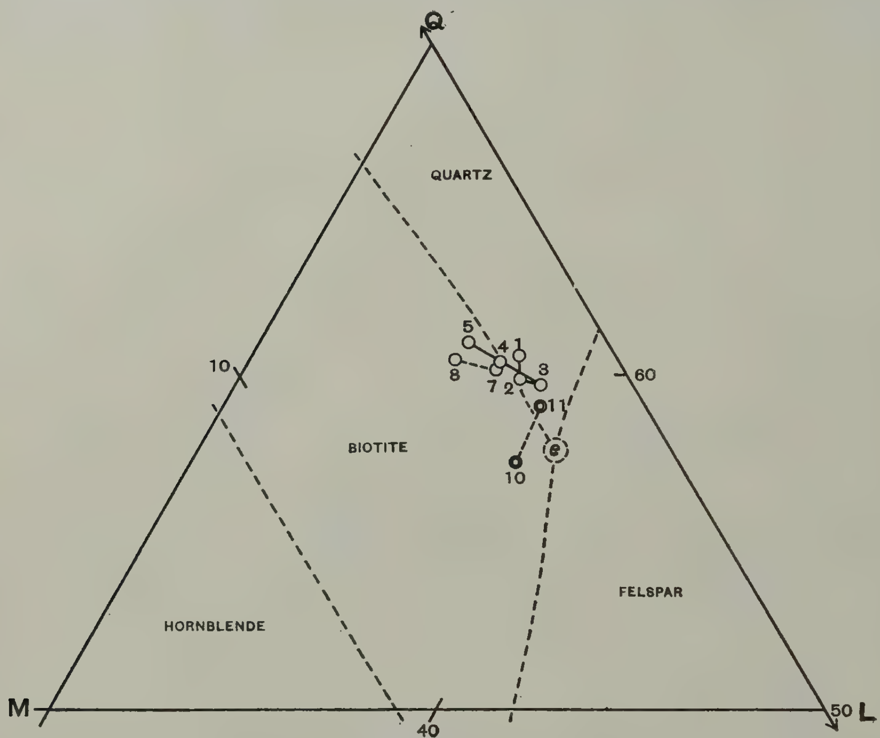

FIG. 11.

In dyke *a*, which holds the partly incorporated xenolith of hybrid microgranite, the end product of the *in situ* differentiation, namely the granophyre (Table I 8) is texturally intermediate between samples 3 and 4 of dyke *f* owing to its more rapid consolidation. The slight difference in chemical composition between the texturally similar rock types of dykes *a* and *f* may perhaps be ascribed to slight contamination of the dyke magma by hybrid granitic material during its upward passage.

A consideration of diagrams 10 and 11 reveals the slightly more basic character of dyke *a*, but otherwise the general trend of the differentiation is similar to that of the chilled portion of dyke *f*. Chemically the principal difference pertains to the slightly more calcic character of a smaller amount of normative felspar and a slightly greater concentration of femic constituents. The higher percentage of biotite and the fact that the most basic and highly zonal type of plagioclase was found to characterise the rock comprising dyke *a*, serve to substantiate this conclusion.

The younger Cape Castle granite which D. L. Scholtz (1946) believes to be the coarse-grained facies of the quartz porphyry masses south of North-West Bay (Table I, Nos. 10 and 11), though differing slightly in composition from the corresponding extreme textural types of the Cape St. Martin's dyke rocks, displays an analogous variation in the alkali enrichment of the normative felspar but in the femic minerals the position is reversed.

In the *k-mg* diagram the available analyses form an elongated zone in Niggli's alkali field. In the k-π ternary diagram, in which the ratio π represents the amount of anorthite to the total felspar and k the ratio of potash to soda-felspar, these analyses fall well within the pacific suite, and occupy the lower half of the field represented by the Western Cape granites, with the exception of Number 5 which lies slightly beyond the field.

Though no phosphorus pentoxide is returned by the analyses 1 to 5, it should be pointed out that a few crystals of apatite were observed in some of the thin sections of the corresponding rocks. The average titania content of the Western Province granites is $0 \cdot 35$ per cent, but rises to $0 \cdot 51$ per cent in the obviously contaminated rock types. In this respect the abnormally low TiO_2 content of the suite of samples taken across dyke *f* and the slightly higher values returned by the two samples from dyke *a* are of interest. Sulphur, fluorine and chlorine are present in normal amounts in all the samples analysed. It will be observed that the outer chilled zone of dyke *f* has a slightly higher fluorine content than the more coarsely crystalline core, while contamination by sea water may be responsible for the higher Cl_2 values of samples 5 and 7 which were taken nearest to low water mark.

The dyke rocks of the peninsula are more acid than the Western Cape granites and show a closer correspondence in chemical composition to the George granites. If the Niggli values (Table I*b*) *si*, *al*, *alk*, *c* and *fm* are used and are graphically plotted with *si* as abscissa in the calc-alkaline differentiation diagram (not represented in this paper), they all fall beyond Niggli's aplogranitic type, which has a *si* of 460. The *al* and *alk* curves fall between those of the George and Cape granites when plotted on C. T. Potgieter's (1950) diagram. The *al* curve is parallel to that of the George granite, while the *alk* curve parallels that of the Cape granites as well as the *alk* curve of Niggli's standard circumpacific differentiation suite curve. The *fm* and *c* curves of these dykes are antipathetic, corresponding respectively to the *fm* and *c* values of the Cape and George granites.

In conclusion, the differentiation trend exhibited by dyke *f* can only be attributed to crystal fractionation along a thermal gradient modified by undercooling.

V PETROGENESIS

From the structural features discussed in the previous chapters it is evident that the intrusions were not haphazardly emplaced but were closely controlled by the structural outlay of the granites. This is shown by the fact that the dykes persistently paralleled the q-joint direction while the thin apophyses and veinlets exploited the s-joint direction. The mechanics of intrusion can now be considered.

Dyke *k*, which has parallel walls with well defined margins and gradually becomes narrower over a considerable distance, indicates a magmatic intrusion along a joint plane by means of dilation. The principal emplacement mechanism whereby the opening of the fissure was concomitant with the incoming of the magma probably

also operated in the case of the narrow dykes *g* and *i*, but was presumably modified by stoping, since these two dykes also contain inclusions of the country rock (Fig. 4). It is clear that in some of the larger dykes the pronounced irregularities in the configuration of the opposite walls cannot be accounted for by dilation alone. It is suggested that block and piecemeal stoping, therefore, also played a part in the process of emplacement. This conclusion is corroborated by the presence of small angular and rounded inclusions of wall rock within the substance of the dykes and the remains of a marginal shatter zone (Plate II, ph. 7) which no doubt owe their preservation to the fortuitous congealing of the dyke magma. In this respect it is significant that, whereas partly detached or slightly displaced inclusions possess angular outlines which match the configuration of the adjoining fragments and contiguous wall rock, the xenoliths which appear along smooth and sinuous contacts possess ill defined margins rounded by spalling and slight magmatic corrosion as a result of a longer period of immersion.

The emplacement mechanism as visualised by the writer is diagrammatically illustrated in Fig. 12.

The granite basement subjected to regional stress induces the opening up of through-going q-joint fissures which are simultaneously infilled by the rising dyke magma (Fig. 12A). Example, dyke *k*.

The dykelet increases in width by further dilation, and differential heating induces minor fracturing of the wall rock. Veinlets searching out these newly formed irregular and diagonal cracks gives the magma access to adjoining master q-joint fractures, which are selectively exploited (Fig. 12B). This stage is illustrated in the field by dyke *i* and its twenty-five feet long apophysal dykelet.

FIG. 12.

With further dilation the advancing apophysal dykelets and veins eventually meet, envelop and wedge off parallel-sided blocks of the country rock. Differential stress may induce the shattering of some of the dislodged blocks and the wall rock in the vicinity of intersecting tongues of magma exploiting the s-joints and adjoining fractures. Dyke *q* and *l* respectively serve to illustrate the initial stage in blockstoping and the formation of embayments. Bifurcating extremities of dykelets initiate swelling and pinching, which when further exploited, may lead to the production of embayments and protuberances of the wall rock as in dykes *c* and *d* (Fig. 12C).

The increase in the width of the adjacent dykes by dilation, and the removal and subsidence of dislodged blocks (Fig. 3) by stoping, now enable the magma to attack the intervening and weakly supported screen of country rock from two sides. This stage in the process of emplacement is illustrated by the closely spaced dykes *d*, *e*, *f*, *h* and *l*, *m*, *n*. The thirteen-feet-long inclusion (Plate I and Plate II, ph. 2) along the northern margin of dyke *m* in all probability represents part of a disrupted septum (Fig. 12D). A stage of more intense stoping activity is initiated and may be accompanied by the foundering of large sections of the country rock screen which have been undermined by magma in depth. The walls of the widening intrusion are also subjected to more rapid but differential attack and the body of the intrusion is cleared by the foundering and settling of all dislodged blocks. This is accompanied by the rapid dissipation of magmatic heat, and consolidation of the magma commences in the less favourably situated and narrower parts of the intrusion, which may be subjected to auto-injection (Fig. 12E and dyke *m*).

On complete consolidation the intrusion, now practically devoid of xenoliths, is characterised by dissimilar though sub-parallel walls. Attention should also be drawn to the fact that although no vertical or horizontal movement of walls is assumed, the amount of offsetting of pre-dyke features in the country rock is not proportional to the width of the intrusion (Fig. 12F). M. P. Billings (1925) described a dyke at Medford showing neither offsets of pre-dyke features nor the effects of differential wall movement. Its emplacement cannot, therefore, be explained by dilation. He believes that its emplacement can be adequately accounted for by his hypothesis of large block stoping.

It is possible that the isolated and apparently plug-like body which severs an otherwise continuous aplite dyke in the wall rock some forty feet south of dyke *r* may be regarded as a very small scale example of intrusion by stoping only.

G. E. Goodspeed (1940, 1952) and B. C. King (1948) ascribed great importance to offsetting as a means of distinguishing between dykes and replacement dykes. According to Goodspeed (1940, p. 189), "one of the best criteria for distinguishing dilation from replacement dykes is the structural evidence of offset or displacement of any distinctive feature such as an earlier dyke, vein or layer in the wall rock". In a recent paper A. K. Wells and A. C. Bishop (1954) conclude that offsetting constitutes an unreliable criterion for distinguishing modes of dyke emplacement if the non-displacement of pre-dyke structures in the wall rock is considered as a three-dimensional problem.

The offset of the contact between the even-grained and porphyritic granite comprising the wall rock of dyke *r* on the Cape St. Martin peninsula serves to illustrate this point, since the observed phenomena can equally satisfactorily be explained by dilation and simultaneous horizontal displacement of its walls or by block and piecemeal stoping. Moreover, further consideration of G. E. Goodspeed's (1940) field and laboratory criteria for distinguishing between dilation and replacement dykes is most illuminating. Of the sixteen listed diagnostic characteristics of the replacement dykes, only three, Numbers 4 and 7 in the first list and number 5 of the second list, appear to be convincing, the remainder being either unsatisfactory or of doubtful significance.

One or more of the following processes have been invoked to account for the origin of dykes infilled with eruptive rock or material resembling igneous rock.

Magmatic dykes	Dilation dykes	1. Forcible intrusion along structurally weak planes.
		2. Passive intrusion along fissures opened by tensional forces, with or without stoping.
	Stope dykes	3. Stoping of large blocks paralleling the initial dilation fissure.
		4. Piecemeal stoping of an initial dilation fissure.
Metasomatic dykes		5. Replacement of the country rock by active migrating solutions.
		6. Mobilised replacement material (Goodspeed, 1952).

It is generally agreed that magmatic dykes owe their origin to the intrusion of magma along structurally weak planes in the country rock, either by forcing the walls aside or by magmatic invasion concomitant with opening and widening of a fissure by tensional forces. As far as the intrusions of the Cape St. Martin peninsula are concerned there appears to be no evidence at all to suggest the metasomatic origin of any of the dykes.

It has been shown that the intrusive magma was highly acid and partly crystalline and hence a very favourable medium for the preservation of any directional fabric structure that may have been impressed upon it at the time of consolidation. The complete absence of any steeply dipping pronounced primary structure and the lack of evidence of the exploitation of sub-horizontal joint systems of the country rock by the intrusive, force one to conclude that magmatic pressure could only have played a very insignificant part during the emplacement of the dykes.

The fact that the trend of the dykes parallels that of the q-joints, which is also the direction exploited by offshoots and dykelets, while the tighter s-joint direction of the country rock is followed by vein-like apophyses which rarely exceed 3 inches in width, strongly suggests the passive emplacement of the dyke magma along tension fissures. This conclusion is further strengthened by the presence of the sharp tapering septa of country rock and the emplacement of the second intrusion in dyke *m*. The later intrusion shows distinct parallelism to the strike and dip of this dyke, but the fading out of its southern contact seems to indicate that the succeeding intrusion must have followed the parent intrusion almost immediately.

It is significant that this dyke, which is the only multiple intrusion on the peninsula, is also the widest and most arcuate intrusion. The wedge-like younger component may, therefore, represent part of the yet mobile fraction of the parent intrusion which invaded an almost completely consolidated portion of its own body as a result of the upwelling of magma induced by the subsidence of a thin unsupported septum of country rock, or alternatively it may owe its origin to the infilling of a wedge-like fissure produced by regional tension on a partly congealed and therefore still weak, curved intrusion.

Although microscopic films of dyke material exploiting joint planes, grain boundaries and felspar cleavages occur commonly in the wall rock, only one other example of auto-injection, this time on a microscopic scale (Plate VI, pm. 6), was observed in the intrusive near the north-western contact of dyke *a*.

It follows that in event of complete crystallisation the microscopic injections and infiltrations into the wall rock exploiting the grain boundaries may provide an interlocking grain relationship or sutured contact between country rock and the dyke substance. Since this phenomenon has commonly been regarded as providing proof of the replacement origin of the dyke material in the past it is essential to draw attention to this argument.

The differential chilling effects and the corresponding variable gradation are striking when the different dykes are compared. The highly variable widths of the chilled margins, even along a single dyke wall, can clearly be ascribed to the width of the dyke and the effects of stoping which induces differential heating of the walls revealed by the contacts along embayments and protuberances. Such variations in the extent and width of chill zones are based on observations on a horizontal plane. Theoretically it is, therefore, to be expected that in a vertical direction the effects of chilling should be even more pronounced. If it were possible to examine outcrops of the same intrusion on a plane a few hundred feet higher than the present one, only quartz porphyries would be present. Similarly a lower datum would probably be characterised by wider dykes of granite porphyry and granite rather than their more highly chilled equivalents. Attention should also be drawn to the fact that the dykes of the south-western and northern parts of the peninsula converge in dip towards the base of the peninsula, which suggests the presence of a simpler and possibly single intrusive body. Accordingly, this textural gradation (so well displayed by the Cape St. Martin intrusives) which may be regarded by some as being of an exceptional or fortuitous character, is in reality a highly instructive exposure at a critical level. D. L. Scholtz (1946, pp. xlvii-lii) has shown that a similar relation occurs in the granite batholiths. He says: "they (the fine-grained suite of granites) appear to attain their maximum thickness in the vicinity of the summits of the plutons and thin out downwards along the sides of the intrusions. In the relatively slightly denuded plutons the fine-grained granite suite may cover extensive tracts such as Paardeberg, but in deeply eroded plutons a normal (chilled) marginal zone may not be present. . . . Extensively eroded plutons may exhibit comparatively broad belts of migmatite passing over into granite towards the contact, and into hornstone away from it, a feature so well illustrated by the Sea Point Contact. . . . The coarsely porphyritic variety of granite has also been observed to grade into the coarsely porphyritic granite porphyry and quartz porphyries by a gradual reduction in the crystallinity of the matrix, the sizes of the phenocrysts remaining approximately constant as in the Stellenbosch and Wellington areas. This phenomenon is strongly suggestive of marginal chilling effects".

The Stellenbosch chill zone referred to above is exposed on the foothills of the Banhoek Mountains near Simondium. A vertical range of fully 1,300 feet can be studied, but because the continuity of outcrops in the vicinity of the contact is interrupted by a heavy overburden, the mapping of the area was relinquished in favour of the well exposed rocks of the Cape St. Martin peninsula. It is noteworthy that the potash felspar phenocrysts measuring about 4×3 cms. were not only scattered throughout both rock types but also distinctly tended to aggregate in irregular clusters on either side of the inferred transition zone at an elevation of 1,650 feet above sea level. At higher elevations the quartz porphyries are characterised by microscopic intergrowths, and from 1,800 feet upwards the quartz porphyry develops a finer grained matrix but still retains the large felspar phenocrysts, while at an elevation of 2,100 feet there appears a completely devitrified glassy rock with well developed perlitic cracks.

The degree of crystallinity, granularity and fabric of chilled margin of an intrusive may provide clues to the crystallisation sequence of the minerals, the concentration of volatiles, flowage of the magma, high- and low-temperature felspars, heat conductivity of the country rock, mechanics of intrusion and cooling velocity. In a completely crystalline rock, however, we are faced with the ultimate or end product which discloses few if any of the intermediate steps. It is, therefore, natural that owing to the lack of evidence the genesis of such an end product may be plausibly

interpreted by many different theories or even from diametrically opposite points of view.

In the chill phases of the rocks under consideration, minerals like zircon, allanite and apatite were the first to crystallise and they were followed by biotite and quartz, and then by plagioclase and microcline-microperthite. The postulated sequence of crystallisation illustrated in Fig. 13 is based on the inclusion-relationship existing between the microscopically determinable minerals occurring in the highly chilled selvages, and on the continuously changing fabric pattern displayed by an undisturbed and gradational succession of textural rock facies between the margin and the core of the intrusion.

The fact that the biotite which envelops the accessory minerals is often seen

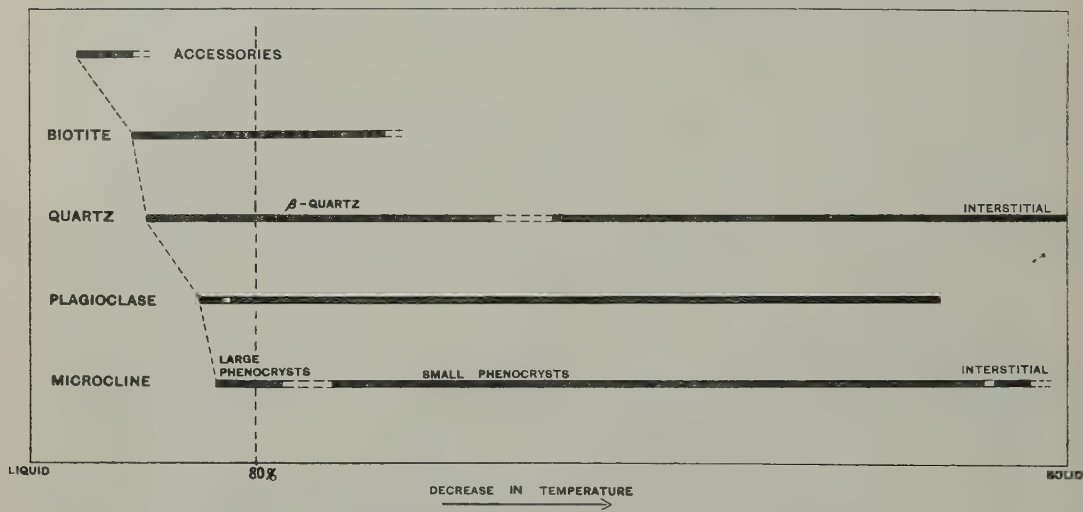

FIG. 13. Diagrammatic representation of the probable order of crystallisation of dyke magma.

to be included by phenocrysts of β-quartz, plagioclase and microcline indicates that it must have commenced crystallisation simultaneously with, or even slightly earlier than, the above-mentioned insets in the chill phase. Inclusions of plagioclase up to 5 mm. in length frequently occur in phenocrysts of microcline but the reverse relation, though observed, is very uncommon. That is, at a stage when at least 80 per cent of the magma was still in a fluid condition (Table II, mode 7) crystals of all the constituent minerals of the rock were already present. This early overlap in the order of crystallisation of the component minerals must necessarily have resulted in the occasional inclusion of one mineral by another. It therefore follows that the application of Rosenbuch's first rule in determining the sequence of crystallisation in phanerocrystalline rocks cannot but fail to yield contradictory conclusions.

Considerable attention has been focussed recently on the presence and physical characteristics of the high-temperature varieties of quartz, plagioclase and alkali felspars such as sanidine and anorthoclase, in hypabyssal and volcanic rocks, as well as the low-temperature forms of these minerals which appear to be restricted to the plutonites.

The absence of high-temperature minerals in plutonic rocks led the transformationists to revive temporarily the heated controversy concerning the genesis of granitic rocks. O. F. Tuttle (1952) and D. L. Reynolds (1952) thoroughly dealt with this question of contrasting mineralogy, but arrived at different conclusions. In the Cape St. Martin rocks the phenocrysts of plagioclase, and xenocrysts of the same mineral derived from the granitic wall rock, occur in the chilled selvages of the dykes. Microscopic investigation proved these minerals to be low-temperature plagioclases ranging in composition from albite to oligoclase. The existence of felspars transitional between the high- and low-temperature varieties is suggested, however, by the optical properties of plagioclases of slightly higher anorthite content. The laboratory work of N. L. Bowen and O. F. Tuttle (1950) seems to indicate that the albite-analbite inversion may be near 700°C, while J. R. Goldsmith and F. Laves (1954*a*) found that the microcline-sanidine inversion occurs at 525°C under hydrothermal conditions. The microscopical evidence which suggests direct crystallisation of microcline and low-temperature plagioclase in the dyke rocks of the peninsula, considered together with the findings of these scientists, serve to indicate that the crystallisation of the intrusive magma occurred at lower temperatures and higher pressures under conditions characterising the plutonic environment whereas effusive rocks normally congeal at higher temperatures and under low pressures.

According to magmatists, the rocks of the rhyolite-granite clan represent the consolidated products of a molten acid magma. Apart from certain paragneisses of metamorphic origin, granite is believed to originate from the slow cooling of an acid magma in depth or from the *in situ* granitisation of enveloping crustal rocks of appropriate composition, on a scale proportional to the volume of the intrusive magma. The more rapidly cooled hypabyssal and volcanic equivalents of granite congealed to form the chemically related acid porphyries and rhyolites.

Both wet and dry transformationists agree that whereas rhyolitic lavas consolidated from an *acid melt*, granite massifs of batholithic dimensions are *not* to be regarded as the ultimate consolidated products of such an acid melt or granite magma as visualised by the magmatists.

The transformationists now no longer regard the intrusive contact as providing proof of the original molten condition of granite plutons (R. Perrin, 1954), and some have even gone so far as to assume the flow and fractional crystallisation of a product of metasomatic origin (G. E. Goodspeed, 1952). In order to account satisfactorily for the observed phenomena characterising the intrusive contact, either mobilisation of the granitic material is assumed, or alternatively, the granitic substance makes its way in as nearly solid masses, as suggested by H. H. Read (1951) in the case of the Cape batholiths, since this granite is not of "granitisation origin at its present level".

If we are to believe the transformationists the only conclusive field evidence for the former existence of an acid melt or granite magma is the chilled contact, since it is unlikely to characterise granite plutons crystallising at depth, that is, the normal batholithic environment.

Singularly, both wet and dry transformationists make a clear statement in the case of rhyolites and their pre-Tertiary equivalents, the quartz porphyries. In his classification of rocks H. H. Read (1943, 1944) recognises three principal groups: the Neptunic, Volcanic and Plutonic; on genetic grounds he separates the granites from the quartz porphyries and rhyolites. In his volcanic division he includes the last named acid effusives and lavas derived from basic magmas. The views of the dry transformationists are illuminating. R. Perrin (1954) believes in the existence of

rhyolitic melt though it may be derived from rootless magmatic sources. D. L. Reynolds (1952) maintains that "there is and can be no evidence for granite magma (as distinct from rhyolite magma) other than the existence and relationships of granite; granite magma is and always has been a purely hypothetical concept". It is therefore clear that the geological evidence concerning the existence of rhyolitic magma is held to be irrefutable even by the most extreme of petrologists. Since no one is willing to dispute the fact that the rhyolite-granite clan consists of texturally different rocks of almost identical chemical composition, it follows that a continuous gradational relation, such as is observed in the rocks of the Cape St. Martin peninsula, can only lead to one conclusion: that rhyolite magma is synonymous with granite magma. That this is not merely a local phenomenon, is clearly borne out by the inferred continuity of the discontinuous outcrop of quartz porphyry—granite porphyry—granite apophysis west of Paarlberg pluton, the quartz porphyry and devitrified glassy chill facies of the granites at Wellington and Simondium respectively, as well as the widespread fine-grained chilled hood facies characterising so many of the coarsely crystalline granite batholiths of the Cape Province.

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GEOLOGICAL MAP
OF THE
CAPE ST. MARTIN
PENINSULA
LEGEND
SHELL (SH) AND SANDY (S) BEACHES
PEBBLE AND BOULDER BEACHES
ACID DYKELETS, VEINS AND APOPHYSES
PORPHYRITIC ACID DYKES
APLITE
MEDIUM TO FINE-GRAINED GRANITE
COARSE PORPHYRITIC GRANITE
LINEATION
ATLANTIC
OCEAN
SH
S
N
100 50 0 100 200 300 400
SCALE IN ENGLISH FEET

PLATE I

PLATE II

Ph. 1 Irregular hornstone inclusion with sharply defined contacts in the older coarsely porphyritic granite.

Ph. 2 Sharply defined and angular granite xenoliths in dyke *m*. Note parallel orientation of phenocrysts along margin of the inclusion in the lower part of the photograph.

Ph. 3 Hybrid microgranite xenolith in the porphyritic granite dyke south of the peninusla. Note the irregular indefinite and hazy boundaries of this partially incorporated xenolith.

Ph. 4 Partially assimilated hornstone xenolith in the coarsely prophyritic old granite.

Ph. 5 Hybrid microgranite xenolith within the granophyric interior of dyke *a*. Note the indefinite penetration of the inclusion by dyke material in the top righthand corner of the photograph

Ph. 6 Isolated and partially assimilated hornstone xenolith completely enveloped by hybrid granitic material in the coarsely porphyritic granite of the Saldanha batholith. Note the presence of the large felspar phenocrysts across the junction of the granite and hybrid granitic material in the top righthand corner of the photograph.

Ph. 7 Shatter zone along the southern contact of dyke *h*.

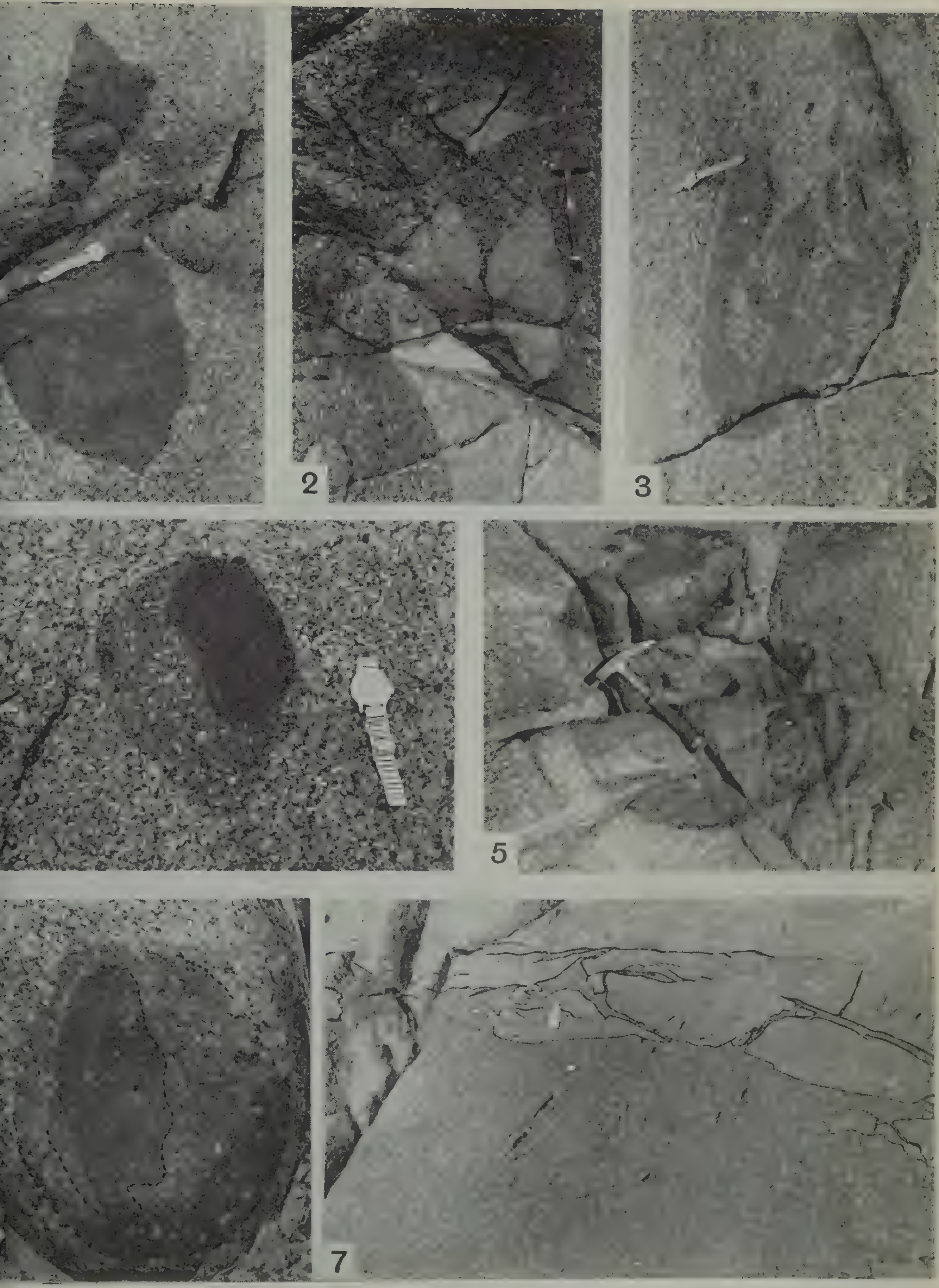


PLATE II

PLATE III

Ph. 1 Southern contact of dyke *h*, illustrating the abundance of large felspar phenocrysts in the coarse rhyolite porphyry as well as vein-like offshoots in the granitic wall rock. Observe that the contact has been displaced in a W.S.W.-E.N.E. direction by a clockwise shear couple.

Ph. 2 The granite porphyry of dyke *e* shows no noticeable chilled selvage and the contact is seen to cut off an aplite vein in the older fine-grained hood facies of the Saldanha batholith.

Ph. 3 Enlarged portion of contact illustrated in ph. 2. Note the absence of any preferred orientation of phenocrysts parallel to the contact.

Ph. 4 Note the sharp contact of dyke *a* with the coarse porphyritic old granite. The chilled selvage of the dyke is distinct and is seen to grade inwards into a rhyolite porphyry with a greater abundance of large insets.

Ph. 5 The rhyolite porphyry contact of dyke *f* showing a few large felspar phenocrysts up to 3 cms. in diameter, and an irregular tapering vein penetrating the surrounding even-grained granite.

Ph. 6 Close up view of the contact seen in ph. 4. Note the sparse distribution of the phenocrysts and the flinty character of the chilled selvage.

Ph. 7 Portion of 350 feet long rhyolite dykelet *k* exploiting q-joint direction of the fine-grained older granite. Note straight and parallel walls of the intrusion, scarcity of phenocrysts and felsitic character of the groundmass.

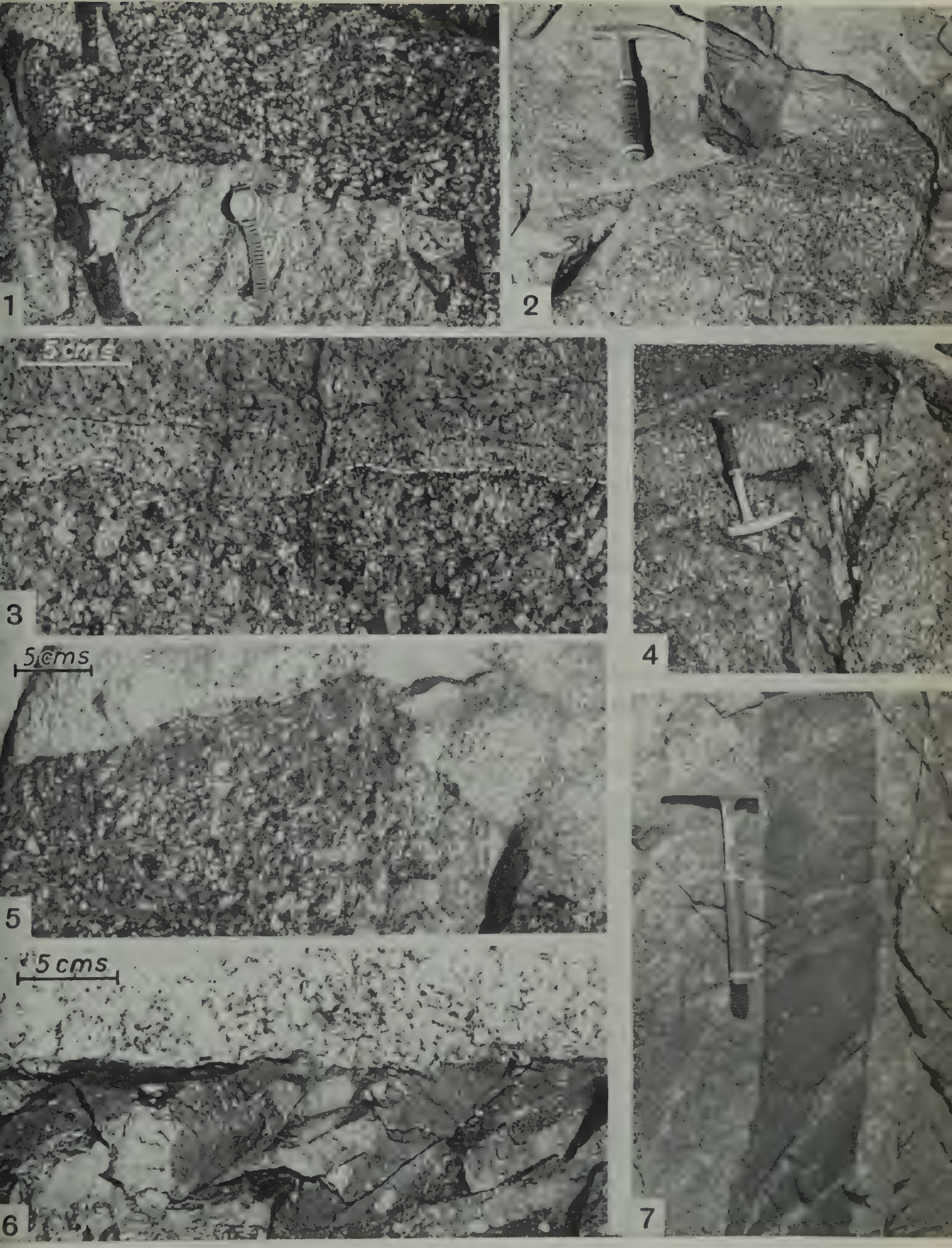


PLATE III

PLATE IV

A. Series of photomicrographs illustrating the textural variation from the contact to the core of dyke *f*. Crossed nicols, × 13.

B. Enlarged photomicrographs of matrix seen in A-series. Ordinary light, × 130.

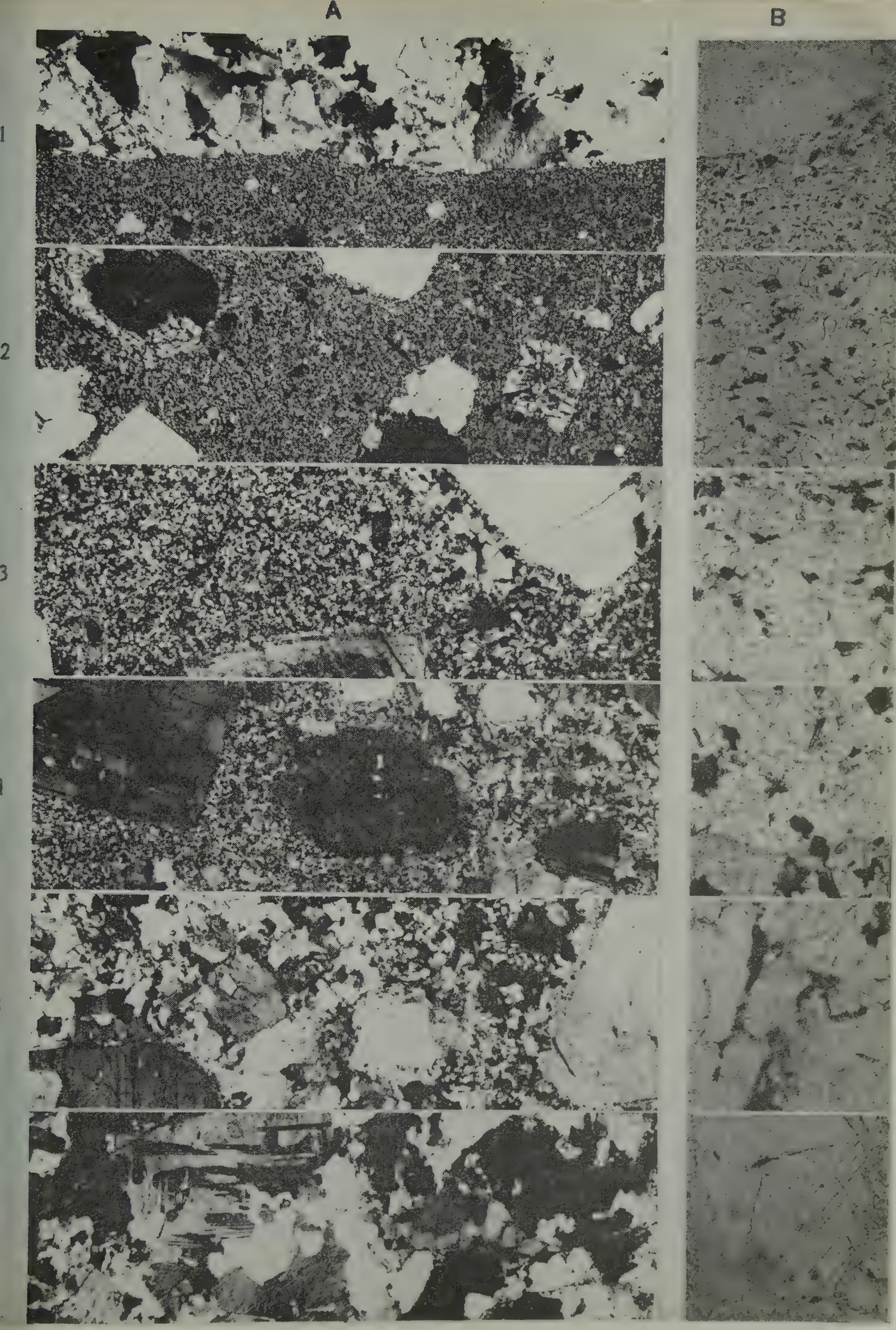


PLATE IV

PLATE V

Pm. 1 Coarse graphic intergrowth of microcline and quartz about strained core of alkali felspar in rhyolite porphyry chill phase of dyke *f*. Crossed nicols, × 44.

Pm. 2 Carlsbad twin of microcline with lower lacinear border and discontinuous upper micro-granophyric collar. Crossed nicols, × 44.

Pm. 3 Uncommon seriate fabric of groundmass of dyke *g* holding only two crystals of microcline showing the rarely developed gridiron twin lamellae or cross-hatching amongst many untwinned microclines (M). Crossed nicols, × 44.

Pm. 4 Granophyric core of dyke *a* showing threads and films of biotite. Ordinary light, × 35.

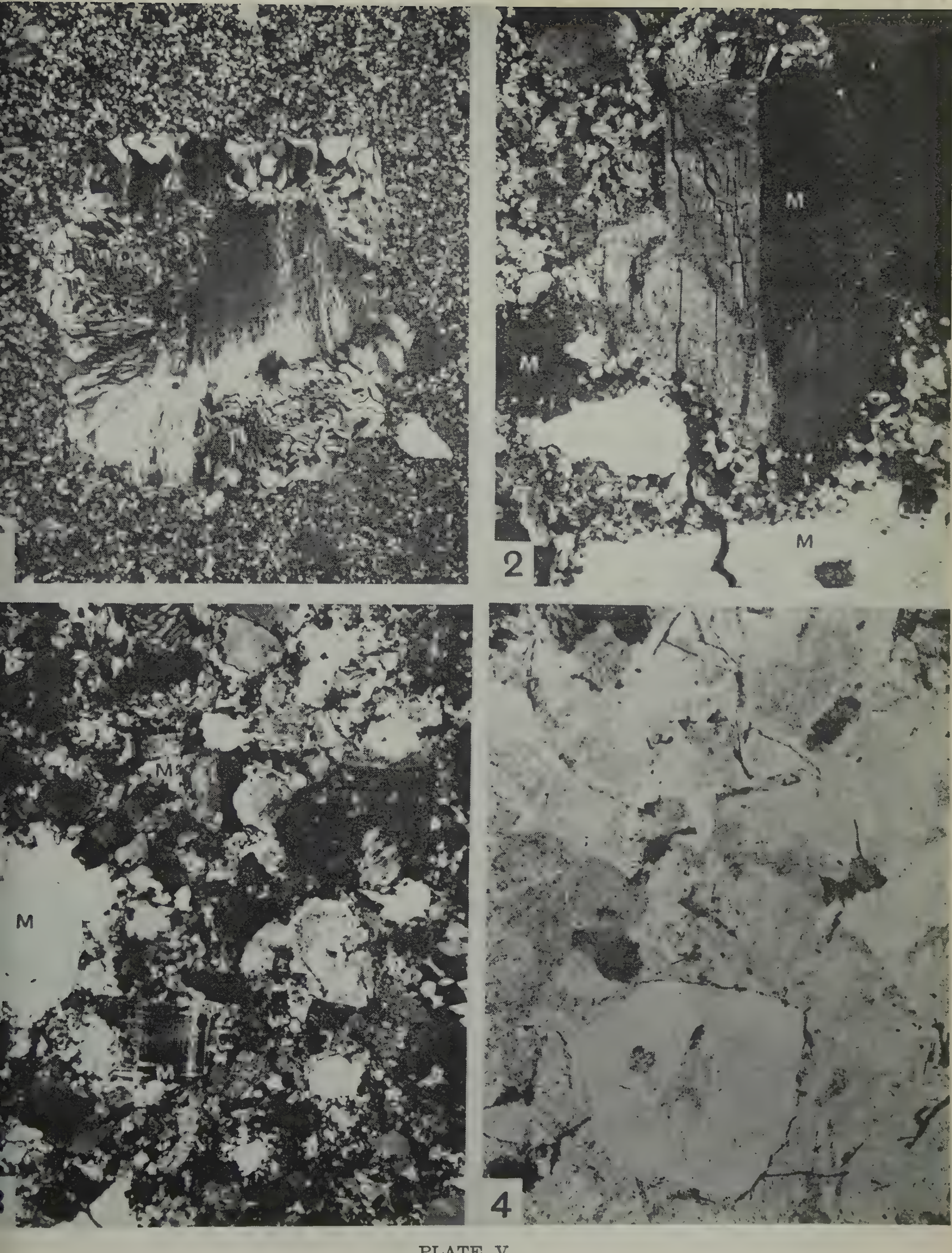


PLATE V

PLATE VI

Pm. 1 Quartz phenocryst exhibiting pronounced resorption, set in a dense matrix holding micro-spherulitic and microgranophyric bodies. Crossed nicols, × 70.

Pm. 2 Idiomorphic β-quartz crystals of different orientation. The slightly resorbed large phenocryst on the left hand side of the photomicrograph has been fractured prior to consolidation of the matrix and is partially re-crystallised along the plane of movement. A re-crystallised phenocryst of quartz is also seen in the righthand side of the photomicrograph. Crossed nicols, ×16.

Pm. 3 Corroded and partially re-crystallised phenocryst of quartz in the granite porphyry facies of dyke *f*. Crossed nicols, × 17.

Pm. 4 The protoshear, which consists of discontinuous biotite film-like aggregates, also shows a minute offset of micrographic growth of quartz in a felspar crystal. Note the distribution of biotite tufts in the groundmass. Ordinary light, × 25.

Pm. 5 Well defined biotite-lined protoshear displacing a fragment of a quartz phenocryst and broadening at one of the two prominent clusters consisting of biotite tufts and scales. Ordinary light, × 17.

Pm. 6 The only example found of micro-auto-injection in the chilled margin of dyke *a*. The biotite tufts and flakes are slightly differently orientated from the adjacent groundmass in the tongue-like injection, and their darker appearance is due to the finer grained groundmass. Crossed nicols, × 10.

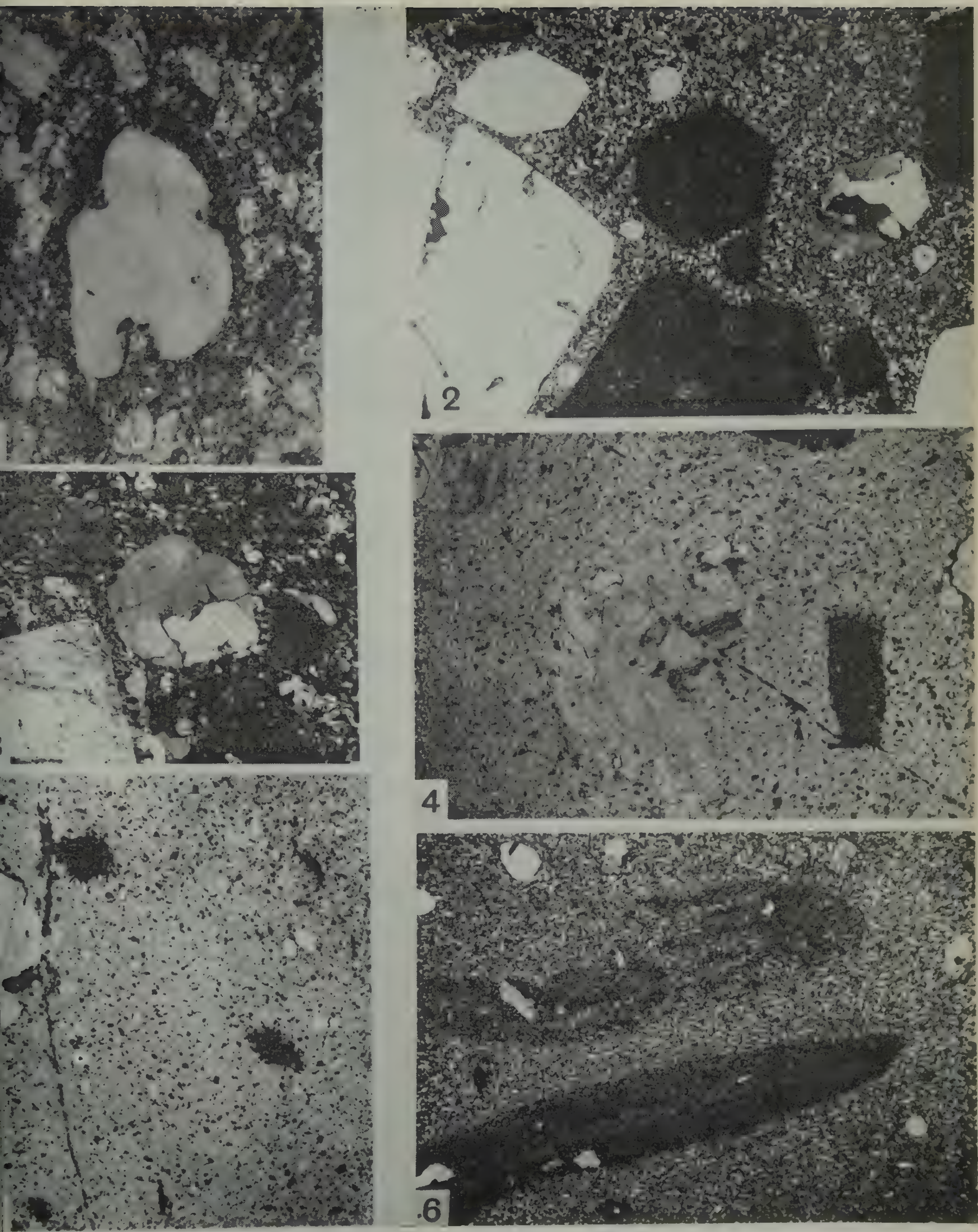


PLATE VI

PLATE VII

Pm. 1 Plagioclase phenocryst with cracked and sericitised core and clear mantle of similar composition. Crossed nicols, × 35.

Pm. 2 Resorption and repair phenomena in a plagioclase phenocryst in the rhyolite porphyry facies of dyke *a*. The pale central portion of the crystal has a lower anorthite content than the partially extinguished portions which are not only of similar composition, but also display analogous zonal structures. Crossed nicols, × 13.

Pm. 3 Euhedral plagioclase ($Ab_{75}An_{25}$) showing irregular cracks and marginal saussuritisation in the chilled selvage of dyke *f*. Crossed nicols, × 35.

Pm. 4 Late aplitic veinlet traversing dyke *h*. Crossed nicols, × 40.

Pm. 5 Intergranular penetration of wall rock by dyke magma along southern contact of dyke *a*. Crossed nicols, × 26.

Pm. 6 Micrographic matrix of rock comprising the central portion of dyke *a*. Crossed nicols, × 40.

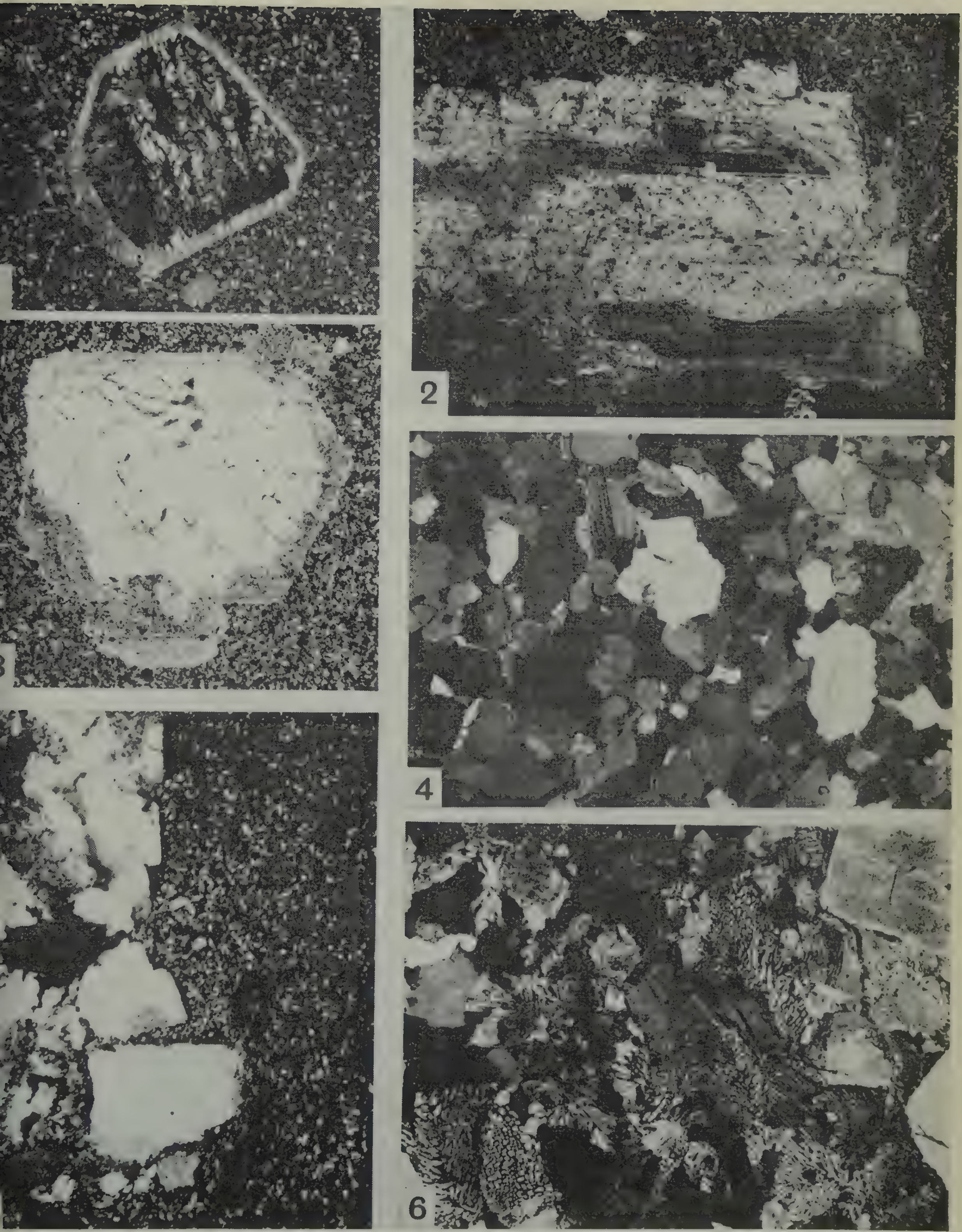


PLATE VII

PLATE VIII

Pm. 1 Concentration of irregular and veinlike perthitic segregations around inclusions in microcline-microperthite phenocryst. Note the delicate intergrowth constituting the sharply defined and discontinuous collar of the phenocryst. Crossed nicols, × 35.

Pm. 2 Preferred orientation of terminally constricted exsolution strings and stringlets in microcline phenocrysts with sieve-like border. Crossed nicols, × 110.

Pm. 3 Zonal crystal of microcline characterised by partial undulose extinction. The zonal structure has been obliterated in the undulant parts of the phenocryst. Crossed nicols, × 10.

Pm. 4 Zonal phenocryst of microcline including small phenocrysts of plagioclase. (N.B. The two halves of the phenocryst are separated by a crack in the section.) Crossed nicols, × 16.

Pm. 5 Pseudo-zoning in a composite microcline phenocryst. True zoning is exhibited by part of the inset in the bottom left hand corner of the photomicrograph. Crossed nicols, × 13.

Pm. 6 Phenocryst of microcline in chilled margin of dyke *p* showing uncommon and irregular resorption. Crossed nicols, × 13.

PLATE VIII

THE GEOLOGY OF THE KLEIN LETABA GOLD MINE IN THE SUTHERLAND RANGE, NORTH-EASTERN TRANSVAAL

By

B. F. WEILERS, B.Sc.

(Submitted December, 1956)

ABSTRACT.

The Klein Letaba Mine as situated within basic schists and amphibolites of the Basement Complex. The mineralised zone consists of nine irregular mineralised shear structures, arranged *en enchelon* from the south-west to the north-east. Mineable ore shoots are subject to a marked structural control, revealed by a gradational structural transition from massive unmineralised amphibolite to well-sheared mineralised quartz-amphibole schists.

Contact metamorphic action, related to the neighbouring granite, resulted in intense metasomatic alteration of the sheared amphibolite host rock, characterised by such minerals as tourmaline, apatite, scheelite, biotite, quartz and acid plagioclase felspar.

The hypothermal gold ores are believed to be closely related to the phase of contact metamorphism. The ores are characterised by ilmenite, lollingite, arsenopyrite, pyrite, pyrrhotite, pentlandite, sphalerite and chalcopyrite. Observations to date suggest that gold was precipitated in the late stages of ore deposition.

This paper was awarded the Corstorphine Medal and the first prize of the Geological Society of South Africa for the year 1956.

CONTENTS

I. INTRODUCTION

The earliest reference to prospecting in the Sutherland Range is given by O. Letcher (1936), who states, "In 1870, Mr. Button, a Natal colonist, and Mr. Sutherland, a Californian miner, explored the eastern districts of the Transvaal, and found gold in the Murchison mountains, one reef being ten feet across. They then descended the Lebombo mountains, and proceeded on foot down the Limpopo, returning with specimens of auriferous quartz, and with the conviction that on the eastern side of the Lebombo mountains, where it intersects the Murchison and Sutherland ranges (lat. 23°50, long. 32°) great deposits of auriferous drift must occur".

Further reference to the area is found in the Mineral Resources (1940): "There is no definite historical record but it is certain that the discovery of gold in this area dates back to the earliest years of European prospecting and gold mining activities in the Transvaal. Thus it is said that the Ellerton and the Eclipse occurrences were both discovered in 1887".

Various attempts at underground exploration at the Klein Letaba Mine (previously known as the Eclipse Mine) have been made, and a minimal tonnage of ore was mined from the No. 4 ore shoot before 1948. No really systematic attempt to develop ore was made until 1948, when the O'okiep Copper Company Limited began an intensive surface diamond drilling programme. On completion of this exploration phase the existing shaft on the property was deepened to the 250-foot level and extensive development of this level by drifting and underground diamond drilling was carried out during the period 1951-1953.

The gold-bearing ore deposits in the Sutherland Range have received scant attention in literature. In 1905 Hans Merensky (1905) states, "In conclusion, I should like to mention some occurrences in the Klein-Letaba Goldfields, especially those of the Eclipse, Birthday and Louis-Moore, the last of which is the only one at present developed. There is a mineralised zone on the Eclipse which contains gold-bearing iron and arsenical pyrites in a mass of actinolite and tourmaline".

Location. The Klein Letaba Mine is located at lat. 23°20′, long. 30°35′ in the Sutherland Range, about 50 miles north of the better known Murchison Range. A good road connects the mine with Mooketsi, the nearest railhead, 40 miles to the south-west. The property is situated on Native Trust Land, and falls in the Magisterial district of the Zoutpansberg.

Acknowledgements. The writer wishes to express his indebtedness to the management of the O'okiep Copper Company Limited, for permission to publish the results of this investigation and for a grant facilitating the research carried out at the University of Stellenbosch.

To Prof. D. L. Scholtz for valuable assistance and constant interest in the progress of the laboratory investigations.

To Prof. P. B. Zeeman of the Physics Department for facilities and aid in the spectrochemical investigations.

II. PHYSICAL FEATURES

The North Eastern Transvaal Low Veld enjoys a subtropical climate. The rainy months are from November to March, the average rainfall approaching about 25 inches per year. Spring, summer and autumn are very hot, but the winter weather is very pleasant.

Malaria is endemic in the Low Veld area, and anti-malarial precautions are advocated during the summer. Great strides in combating the disease have been made during the past few years, and many areas have been virtually relieved of it.

Drainage is effected by the Klein Letaba, Tsama, Shingwedzi and Molototsi rivers, which are also the sources of water for domestic purposes in the area.

Geomorphologically the Low Veld surface is an erosion feature developed on the readily weathering ancient crystalline Basement Complex. The surface is gently rolling with a gentle fall towards the east, and stands at an elevation of about 1,000 to 1,500 feet above sea level. The Sutherland Range forms a series of prominent landmarks composed largely of amphibolite and basic schists of the Swaziland System. The Range stands at an elevation which corresponds to the erosion bevel termed by M. S. Taljaard (1936) "the plain of the second order". Development of the Low Veld stage of erosion is regarded by M. S. Taljaard as the 'plain of the third order". These two erosion stages are pre- and post-Cretaceous respectively.

III. REGIONAL GEOLOGY AND STRUCTURE

The Sutherland Range is composed of Old-granite and gneiss invading and assemblage of rocks belonging to the Swaziland System and the Jamestown Intrusive Complex. These archaean basement rocks have been subjected to intense regional and contact metamorphism resulting in a varied assortment of metamorphosed products comprising amphibolites, basic schists, serpentinites, 'banded ironstones' and quartzites.

Geologically this area has received little attention in the past. The only reference to these Swaziland basement rocks made by A. L. du Toit (1939) is that "a minor east-west belt crosses the upper Letaba River."

Paucity of outcrop, combined with a lack of resistant horizons, provides little structural control upon which an adequate regional lithological classification of the basement rocks constituting the Range may be made.

Within the confines of the Sutherland Range, the Swaziland rocks attain a maximum development in the tract of ground between the Ellerton and Giant Reef Mines (see Fig. 1). The rocks here constitute a belt of basic schists, amphibolites and quartzitic sediments trending in an east-northeasterly direction. The schist belt attains a maximum width in the Sudeten-Boltmans Beauty area where it crosses the Tsama River. South of Bend Store this schist belt bifurcates into two prominent limbs separated by a nose of granite-gneiss. Intrusive into the entire rock assemblage on a regional scale are swarms of younger dolerite dykes, which conform to a constant structural pattern.

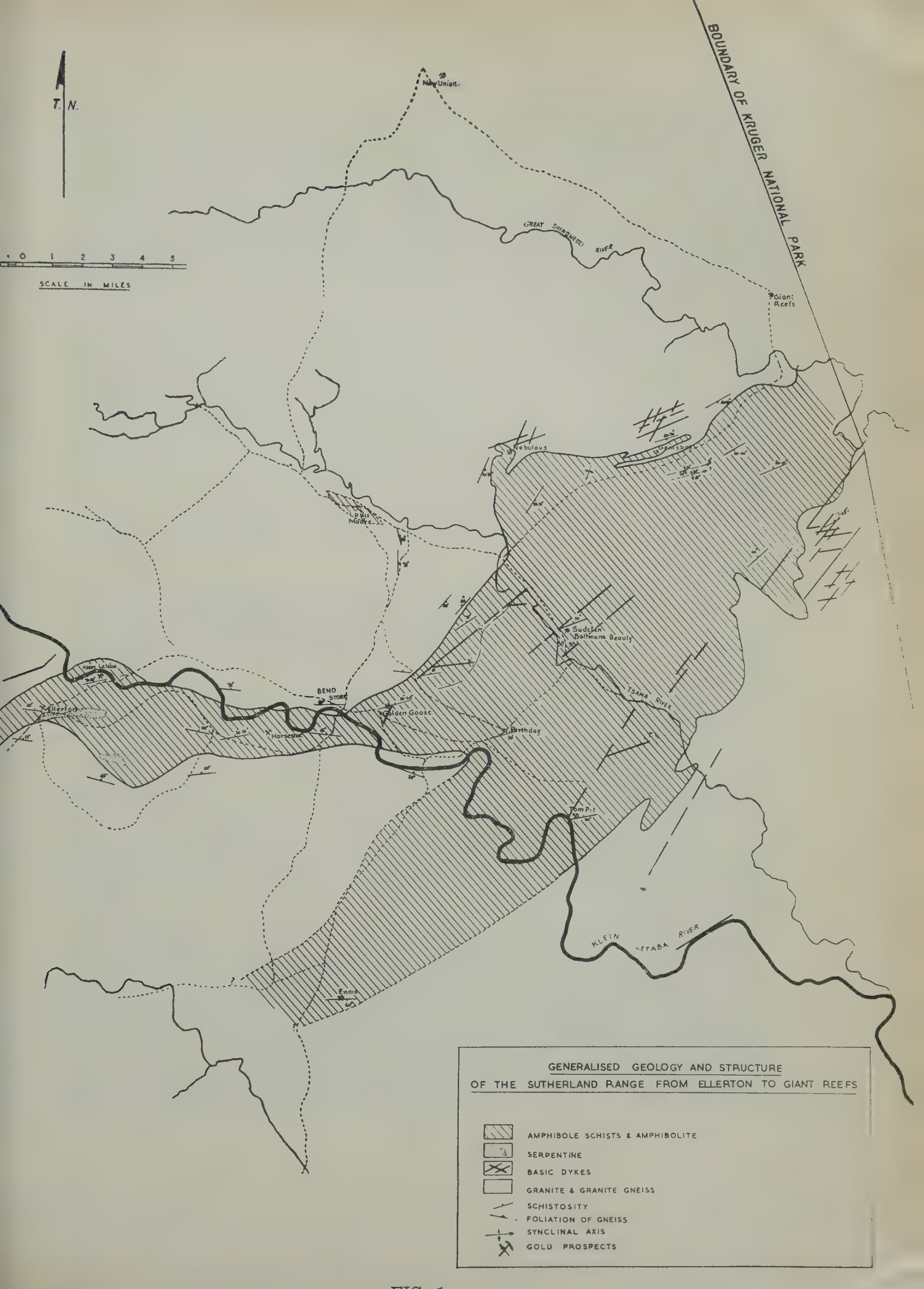

T. N.
SCALE IN MILES
BOUNDARY OF KRUGER NATIONAL PARK
New Union
GREAT SHINGWEDZI RIVER
Giant Reefs
Nebulous
Louis Moore
Sudetan
Bothmans Beauty
TSAMA RIVER
Klein Letaba
Ellerton
BEND STORE
Golden Goose
Horseshoe
Birthday
Tom Pit
KLEIN LETABA RIVER
Ennis
GENERALISED GEOLOGY AND STRUCTURE
OF THE SUTHERLAND RANGE FROM ELLERTON TO GIANT REEFS
AMPHIBOLE SCHISTS & AMPHIBOLITE
SERPENTINE
BASIC DYKES
GRANITE & GRANITE GNEISS
SCHISTOSITY
FOLIATION OF GNEISS
SYNCLINAL AXIS
GOLD PROSPECTS

FIG. 1.

Amphibolites and amphibole schists collectively comprise the major rocks of this belt of intensely metamorphosed rocks. From field evidence the amphibolites may be classified into three distinct locality types, designated the Klein Letaba, Birthday and Louis Moore respectively. The Klein Letaba amphibolites consist of radiating crystal aggregates of actinolite, tremolite and anthophyllite with or without subsidiary talc and chlorite. Around the Birthday Mine extensive outcrops of amphibolite composed of hornblende and plagioclase felspar predominate. This hornblende-rich type of amphibolite weathers into rounded boulders, resembling the spheroidal weathering of certain ultrabasic and basic igneous intrusives. The above two types of amphibolite are considered to be the metamorphosed equivalents of ultrabasic to basic igneous masses, and may be partly correlated with the No. 3 division of the Swaziland System in the Murchison Range (O. R. van Eeden et al., (1939). For the third type of amphibolite, restricted to a narrow lens of basement rocks around the Louis Moore Mine, a different derivation is envisaged by the writer. This amphibolite composed of actinolite, tremolite, forsterite, diopside with minor spinel, phlogopite and calcite, represents a mineral assemblage which, according to Harker (1932), may be derived from the regional metamorphism of an impure magnesian limestone. The association of this amphibolite with mineralised gold-bearing veins in a gangue of calcite and dolomite tends to support this view.

Various types of amphibole schists, among which varieties rich in actinolite, tremolite and talc-chlorite predominate, are widely distributed throughout the schist belt.

The sedimentary members comprising the belt are represented by quartzites, cherts and the conspicuous 'banded ironstones'. The 'banded ironstones' have in all probability developed from the oxidation of siliceous muds. The intensely contorted structure of the ironstones in regions of relatively undisturbed country rocks, points to a deformation of the sediments whilst still in a semi-plastic condition.

Lenticular masses of serpentine, situated south of the Ellerton and Klein Letaba Mines, are considered by Van Zyl and others (1942) to be derived from an original ultra-basic rock type rich in olivine. These ultra-basics may be correlated with the Jamestown Igneous Complex.

The large ubiquitous masses of gneissose granite that so typify the Low Veld generally are widely represented in the area. In appearance and mineralogical composition the granite-gneiss is very similar to that described by A. L. Hall (1912) in the Murchison Range. The granitic-gneisses are invariably accompanied by irregular masses of coarse-grained pegmatite. One such pegmatite at Pala Kop, situated 2 miles due west of the Klein Letaba Mine, has been actively prospected and mined for mica, beryl, tantalite and cassiterite.

Three intrusions of syenite occurring in the area are recorded by J. W. Brandt (1948). The syenites are intrusive into the granite-gneiss and are considered to be a phase of the Palabora Complex.

Structure. The effect of intense regional metamorphism has imparted a pronounced schistosity to the basic schists comprising the Sutherland Range. The major structural features characterising the schist belt between the Ellerton and Giant Reef Mines are outlined in Fig. 1.

The regional schistosity is variable (showing pronounced changes in strike directions) in that the schistosity trends about east-west in the

Ellerton-Bend Store area, N. 40°E. in the Golden Goose-Gemsbok area and again N. 70°E. where the schist belt enters the Kruger National Park. Dips are generally steep to the north, though local structural reversals are not uncommon.

The remarkable parallelism exhibited by the foliation of the granite-gneiss to this curving trend of the schistosity is a striking feature. Attention is drawn to the synclinal structure of the north-east sector of the schist belt, a structure which is exactly duplicated by the neighbouring granite-gneiss.

The schist belt as a whole appears to have been isoclinally folded, but whether this structure forms part of an originally larger anticlinorium or synclinorium cannot be deduced from structural evidence still retained within this archaean formation. Since the direction of flow cleavage or schistosity within the schists may be regarded as parallel to the direction of maximum strain, the flow cleavage could have resulted from a non-rotational stress acting approximately from north to south, or to a rotational stress acting as a couple from some other direction. The former mode of origin is preferred by the writer.

Younger diabase dykes, possibly of Karroo age, form a constant structural pattern composed of two major directional trends striking N. 40°E. and N. 84°E. respectively.

The existence of a dome-shaped structure in the granite to the south of the Zoutpansberg was first recognised by M. S. Taljaard (1936), who believes that this dome forms the floor of the Waterberg System. J. W. Brandt (1948), who bases his evidence largely on the work of F. C. Truter (1945), thinks that the doming of the granite floor originated as a result of compressional forces acting from a southerly direction in the Transvaal Drakensberg and from a northerly direction in the Zoutpansberg in post-Waterberg times. These compressional forces induced the formation of zones of intense compression and tension. The zones of tensional stress are believed to have controlled the intrusion of the syenite magma. Repeated tensional stresses are considered to have increased in post-Karroo times, resulting in the formation of prominent joint systems frequently exploited by dolerite intrusions.

If it is assumed that the granite-gneiss intruded the Swaziland rock assemblage before folding, then abundant evidence of intense dynamic metamorphism should be retained within the gneiss. That no such dynamic deformation is recognised has been repeatedly stressed by A. L. Hall (1920), who states: "Mechanical deformation can certainly be detected under the microscope, but is only feebly expressed and altogether negatives the view that the banding has arisen from the regional metamorphism of an originally massive and rigid granite." O. R. van Eeden (1941) considers that as the gneisses of the Barberton region appear to be elongated parallel to the length of the mountainland, folding and intrusion must have overlapped. Evidence to be presented in a later chapter indicates that at the Klein Letaba Mine pneumatolysis, metasomatism and hypothermal mineralisation related to a granite source are controlled by severe dynamic deformation of the amphibolite host rock. This structural control provides further evidence for regarding the granite-gneiss as a syn-orogenic feature.

IV. MINE GEOLOGY AND STRUCTURE

The Klein Letaba Mine is situated within basic schists and amphibolite of the Basement Complex. This belt of metamorphic rocks which is about 3.5 miles wide at the mine is flanked to the north and south by granite-gneiss. The mineralised ore bodies are located approximately 3,000 feet south of the contact between the basic schists and the granite-gneiss.

Paucity of outcrop necessitated considerable surface trenching to delineate potential mineralised zones on the surface. The mineralised shears, which constitute the gold-bearing ore bodies, weather readily and are concealed by slumped rock in the trenches.

Surface outcrops over the mineralised area are virtually restricted to sporadic lenses of coarse-grained amphibolite and amphibole schist. A bifurcating younger dolerite dyke is located to the north and west of the Main Shaft and has been traced along its strike for 1,600 feet. The dyke which occasionally exceeds 100 feet in width strikes N. 70°E and dips 60-70° to the north-west. Diamond drilling has disclosed a flat-lying apophysis, 10 feet wide, branching off on the footwall side of the main dolerite dyke. This offshoot transects the rock formations and the ore bodies.

The main rock types comprising the mine suite may be classified as follows:

Coarse-Grained Amphibolite. This rock is characterised by a thoroughly haphazard texture. Greenish to grey in hand specimen, it is composed of radiating crystal aggregates of tremolite and actinolite. Subordinate mica is present in a few places. The amphibolite constitutes the host rock to the tabular mineralised shear zones and quartz-sulphide lenses. The main body of coarse-grained amphibolite, as outlined on the section through boreholes KL17 and KL19 (Fig. 3A), is at least 240 feet wide.

Fine-Grained Amphibolite. Irregular lenses of tough fine-grained amphibolite are scattered throughout the mine. A typical lens of this type, outlined by drilling on the 250-foot level, lies 15 feet north-west of the Main Shaft (see Fig. 2). The lenticular bodies rarely exceed 20 feet in width, and pinch out over a vertical range of about 200 feet. Actinolite appears to be the major mineral constituent.

Amphibole Schist and Talc-Amphibole Schist. The amphibole schist comprises both tremolite-rich and actinolite-rich varieties. The degree of schistosity is variable and may range from virtually disorientated amphibolite to strongly schistose types in which talc occasionally becomes a dominant mineral constituent. The colour of the schist varies from grey to greenish, depending on whether tremolite predominates, or actinolite with or without talc. A relationship between the degree of shearing and the quantity of talc was noted in the core-logging.

Quartz-Mica-Amphibole Schist. This rock is a tough, dark-coloured schist, composed of quartz, mica, tourmaline and slender amphibole needles. Schistosity is usually well developed, especially in the mineralised shears. A broad zone of drag-folded quartz-mica-amphibole schist, occasionally garnetiferous, outlined in the section through boreholes KL11 and KL22 (Fig. 3B) attains a maximum width of about 200 feet near the 600-foot level. It is noteworthy, however, that this zone displays a gradational contact relationship to the enveloping coarse-grained amphibolite.

These horizons of intense shear are very important, as they appear to have acted as major loci for the ore-bearing solutions.

Aplite and Pegmatite. Aplite, and rarely pegmatite, form irregular discontinuous sills in the amphibolite. In plan and section it has been established that concordant sheets of aplite lie on the continuation or projected continuation of mineralised shears and quartz-sulphide veins, beyond the point where the mineralised structures pinch out. This feature is well illustrated at the east end of No. 3 Vein on the 250-foot level at co-ordinates Y +258,877, X —207,780, and also on the upward extension of No. 5 Vein in Fig. 3B.

Porphyritic Intrusive. A sill-like body, 1-2 feet wide, has been identified in the underground workings and diamond drill cores. This sheet which may be intrusive, is orientated parallel to the schistosity of the enclosing rocks, and consists of pale coloured phenocrysts in a fine-grained micro-crystalline groundmass.

Structure. The rocks at the Klein Letaba Mine have suffered intense dynamic deformation, with development of a regional east-north-east schistosity, dipping steeply northward. The mineralised zone, having a total length of 800 feet on the surface, consists of nine irregular mineralised shear structures, arranged en echelon from the south-west to the north-east, striking N. 85°E. and dipping 82° to the north. Five of the mineralised loci recognised in the mine form distinct geological structures. These are the so-called Nos. 2, 3, 4, 5 and 9 Veins which are mainly bands of mineralised well-sheared quartz-mica-amphibole schist.

The important No. 2 Shear Vein, which contributes approximately two-thirds of the potential ore delineated by exploration, has a maximum width of 37 feet. It is a complex structure made up of branching and looping shears. The vein has been traced out by development and diamond drilling over a strike length of 470 feet on the 250-foot level.

It is noteworthy that the mineralised and unmineralised shears tend to horse-tail or feather out along the strike. This tendency is well displayed by the No. 4 Vein shear, located 40 feet west of the Main Shaft on the 250-foot level.

Weak slickensiding, pitching at about 45° to the east, can be discerned in a few localities. The ore bodies have the same general pitch. This easterly trending pitch of the mineralised ore shoots is a regional feature of the gold deposits of the Sutherland Range, and is apparently related to regional deformation.

Three well defined sets of joints were mapped, striking N. 55°W., N. 83°E., N. 5°W., respectively. Post-mineralisation movement along the N. 5°W. joints were noted; displacement, however, is seldom more than 0.5 feet.

Although disseminated ore occurs throughout the 200-foot-wide tract of amphibolite and related schists between the No. 2 and No. 9 Veins, mineable ore shoots are subject to a marked structural control. This marked structural control of mineable ore shoots to shear loci is well illustrated in borehole KL19 (see Fig. 4).

Two gold-bearing horizons at Klein Letaba may thus be recognised viz., the mineralised quartz-mica-amphibole schists and massive quartz-sulphide lenses. Some of the quartz-sulphide lenses occur in the mineralised shears, others lie parallel to the shears, close to or at some distance away from the main structures.

Detailed mapping and study of borehole cores reveal a gradational structural transition from massive amphibolite to well-sheared quartz-amphibole schist. It should also be noted that mineralised shears tend to be flanked by, or to grade into, amphibole schist along strike (see Fig. 2).

Depth in Feet	Rock Type	Assay		Width in Feet	Dwt/Ton	Vein Correlation
		From	To			
743·5	Coarse-grained Amphibolite					
747·5	Sheared, mineralised	743·5	747·5	4·0	9·10	No. 5 Vein
753·5	Coarse-grained Amphibolite					
761·0	Sheared, mineralised	753·5	761·0	7·5	20·06	No. 4 Vein
768·5	Coarse-grained Amphibolite					
772·0	Amphibolite with 0.75′ shear	768·5	772·0	3·5	3·90	No. 3 Vein
787·5	Coarse-grained Amphibolite					
795·0	Sheared, mineralised	787·5	800·0	12·5	5·13	No. 2 Vein
800·0	Coarse-grained Amphibolite					

Fig. 4.—Table showing the relation between structure and mineralisation B.H. KL19 — Klein Letaba mine.

V. PETROLOGY AND PETROGENESIS OF THE MINE SUITE

A. Petrology.

The broad macroscopic features of the rock types constituting the mine suite, and their relationships to the overall structural control have been stressed in the preceding chapter. Microscopic investigation clearly indicates a sequence of distinct mineralogical and textural variations, grading from relatively fresh unaltered amphibolite to amphibole schist and finally quartz-mica-amphibole schist, induced by severe shearing stresses and subsequent metasomatic action.

The petrology of the metamorphic rocks and younger intrusives may now be considered.

Amphibolite. The definition by Holmes (1920) of an amphibolite as "a granulose or glomero-blastic metamorphic rock, consisting essentially of amphibole and plagioclase, and often containing quartz, epidote or garnet", is not strictly applicable to the Klein Letaba amphibolite, in that plagioclase felspar, quartz etc. are essentially lacking.

The principal features of the amphibolites studied make possible a classification into two main textural types, namely coarse-grained and fine-grained amphibolite. Mineralogically, however, the two types are essentially similar.

In thin section the coarse-grained amphibolite is characterised by a disorientated pan-xenomorphic texture. This texture is defined by P. Niggli (1954) as follows: "Form development is subject to very considerable mutual interference, so that prevailing shapes are largely foreign to the crystals exhibiting them."

The unaltered rock is composed of hypidiomorphic to xenomorphic fibres and prisms of actinolite and tremolite, with a prominent (110) cleavage. The grain size is invariably coarse and fibres up to 1 cm. in length are not uncommon. Actinolite predominates over tremolite.

Actinolite is pale green in colour with faint pleochroism, having $\alpha =$ colourless, $\beta =$ greenish yellow, $\gamma =$ pale green, and an absorbtion formula $\gamma > \beta > \alpha$. Further determinations by means of the U. M. stage yielded the following optical constants:

$2V_\alpha = 79°$ $\gamma/c = 18°$.

The following refractive indices were determined by the immersion method:

$\alpha = 1.628\ (\pm)\ 0.003$ $\gamma = 1.650\ (\pm)\ 0.003$.

Birefringence $(\gamma - \alpha) = 0.022$.

According to A. N. Winchell (1931), the mineral corresponds to a formula of $H_2Ca_2Mg_5Si_8O_{24}$ with 20 per cent $H_2Ca_2Fe_5Si_8O_{24}$.

Colourless tremolite is similar in optical properties to actinolite, but exhibits a smaller extinction angle.

The amphibolite is subject to a marked degree of alteration, and the unaltered rock is rarely encountered in the mine suite. The degree of alteration is variable, being most pronounced in the amphibolite immediately adjacent to the metasomatised shear zones, and may be broadly classified into three distinct zones as follows:

(1) Outer zone (12—30 feet from shear zone).

Laths and fibres of actinolite and tremolite are replaced by micaceous aggregates of pale green chlorite and pale brown phlogopite. Replacement is confined to intergrain amphibole boundaries and prominent 110 cleavage traces. The individual chlorite and phlogopite flakes have a grain size that rarely exceeds a length of 0.17 mm. and 0.40 mm. respectively. Small optic axial angle, straight extinction and very low birefringence with the refractive index of $\gamma = 1.580\ (\pm\ 0.003)$ identify the chlorite as penninite. Accessory small ore grains replace amphibole.

(2) Intermediate Zone (6—12 feet from shear zone).

Acicular aggregates of amphibole are occasionally bent and broken, being extensively replaced by masses of fibrous talc, chlorite and occasionally phlogopite. Scattered ore grains are somewhat more prominent than in the outer zone.

(3) Contact Zone (less than 6.0 feet from shear zone).

Mechanical deformation of amphibole laths is more pronounced, the broken and crushed crystals having partly recrystallised into a dense aggregate of amphibole fibres. Phlogopite (with subsidiary talc and chlorite) becomes a major constituent and may predominate over amphibole, which it replaces

(see Fig. 5). The phlogopite is pale brown in colour and is distinctly pleochroic, with the following optical constants:

$2V_\alpha \pm 1$—3°, straight extinction, prominent (001) cleavage, $\gamma = 1.600$ ($\pm$ 0.003).

Fig. 5. Large actinolite crystal showing pronounced replacement by pale coloured phlogopite. Note how replacement is controlled by prominent (110) cleavage of the amphibole (× 375). Actinolite (white), phlogopite (stippled.)

Calcite replaces amphibole, and the size and amount of ore grains notably increases. Scattered idioblastic prisms of pale green pleochroic tourmaline with absorption $w > \varepsilon$, replace actinolite and tremolite along (110) cleavages. Anthophyllite occurs as occasional slender fibres with the following optical properties:

$2V_\gamma = 79°$ $\gamma/c = 0°$
$\alpha = 1.612$ ($\pm$ 0.003), $(\gamma - \alpha) = 0.019$.
$\gamma = 1.631$ ($\pm$ 0.003).

According to A. N. Winchell (1931) the composition corresponds to mag-anthophyllite with 15% molecular ferro-anthophyllite.

A specimen of well mineralised amphibolite situated within the No. 3 Shear Vein (at co-ordinates Y +258,807; X —207,204 (see Fig. 2) is worthy of mention. The texture is pan-xenomorphic and actinolite is the dominant amphibole. Hypidiomorphic to xenomorphic aggregates of tourmaline (schorlite), apatite, scheelite and minor biotite replace amphibole (see Figs. 6 and 7), and are in turn replaced by sulphides. The buff to olive brown schorlite is strongly pleochroic with absorbtion $w > \varepsilon$. The mineral is uniaxial negative

Fig. 6. Actinolite crystals extensively replaced by apatite and scheelite, mainly along intergrain boundaries (× 600). Actinolite (stippled), apatite (white), scheelite (dotted).

Fig. 7. Biotite invades actinolite along a crack and prismatic cleavage traces. Tourmaline and sulphides are also seen to invade the actinolite along its cleavage traces (× 625). Actinolite (stippled), tourmaline (dotted), sulphides (black), biotite (ruled).

and the following refractive indices were determined by the immersion method:

$\varepsilon = 1.629$ ($\pm$ 0.003), $w = 1.652$ ($\pm$ 0.003), $(w-\varepsilon) = 0.023$. Apatite occurs in close association with scheelite. The former mineral is uniaxial negative, and the following refractive indices were determined:

$\varepsilon = 1.641$ ($\pm$ 0.003), $w = 1.645$ ($\pm$ 0.003), $(w-\varepsilon) = 0.004$. Scheelite is recognised by its uniaxial positive character and high refractive index. Under the iron arc the mineral exhibits a bright blue fluorescence, in accordance with the experimental results quoted by F. R. van Horn (1931). Positive identification of scheelite was obtained by a qualitative spectrochemical analysis and the characteristic tungsten lines 3049.7 A°, 3046.4 A°, 3017.4 A° 2947.0 A°, 2944.4 A°, were readily identified.

Amphibole and Talc-Amphibole Schist. The amphibole schists characterise a mineralogical and textural rock type intermediate between the pan-xenomorphic amphibolite and the strongly sheared quartz-mica-amphibole schist. The degree of metasomatic alteration appears to be directly controlled by the severity of the shearing stress, metasomatism being most intense in those schists which display a pronounced nematoblastic texture.

Actinolite is the dominant constituent of the schists with a feebly developed S-plane structure. The amphibole laths are frequently bent and broken, and have recrystallised into nests of fine-grained radiating aggregates. Phlogopite and penninite replace actinolite. Apatite is occasionally prominent and rarely encloses small grains of scheelite.

In the amphibole schists displaying a moderate nematoblastic texture a weak foliation is developed. Schorlite, with minor quartz and oligoclase felspar constitute the thin 0.50 mm. wide foliae. Calcite may replace actinolite and is in turn replaced by quartz. Where replacement by calcite is prominent, anthophyllite partly substitutes for actinolite. Microscopic fracturing at an angle of 50° to the S-plane structure was noted. The fractures, which are approximately 0.16 mm. wide, off-set the tourmaline-rich foliae. Sulphide grains in a gangue of quartz and albite-oligoclase felspar partly fill the fractures, which are notably devoid of tourmaline.

When the texture is strongly nematoblastic, the schist is always distinctly foliated, the average width of the tourmaline-rich foliae rarely exceeding 5.0 mm. Tourmaline, may now predominate over actinolite and anthophyllite. Quartz and albite-oligoclase felspar become more abundant, and biotite largely substitutes for phlogopite.

The talc-amphibole schists are markedly schistose. Talc with subsidiary chlorite extensively replaces actinolite and/or tremolite, and only vestiges of the latter minerals are still in evidence. The rock is speckled with granular magnetite, and an occasional idioblastic crystal of pale-green tourmaline was noted. The tourmaline is distinctly zonal, consisting of a buff coloured core and a pale green mantle. A few grains of rutile frequently accompany tourmaline.

Mention may here be made of a dark, microcrystalline, antigorite schist encountered in one drill hole on the 250-foot level. The rock is distinctly schistose in thin section, and minute flakes of antigorite constitute the groundmass. Moderate $2V_\alpha$, negative elongation, low birefringence, and an index of $\beta = 1.570$ ($\pm$ 0.003) indicates a mineral composition represented by antigorite with 12 per cent molecular amesite. Streaks and granular

aggregates of magnetite are associated with penninite. A dusting of small calcite grains occur scattered throughout the groundmass.

Quartz-Mica-Amphibole Schist. Important as loci for ore-bearing solutions, these schists characterise the zones of most intense shearing and subsequent metasomatic action. Extensive replacement of amphibole by biotite and quartz finds maximum expression in these schists. The nature of this replacement is illustrated in Fig. 8. Pronounced lepidoblastic texture and foliation is always prominently developed. The typical schist is composed of biotite, tourmaline (schorlite) and quartz with subsidiary plagioclase felspar and vestiges of unreplaced amphibole. Disseminated grains of sulphide replace the silicates, and are orientated parallel to the schistosity (see Fig. 9).

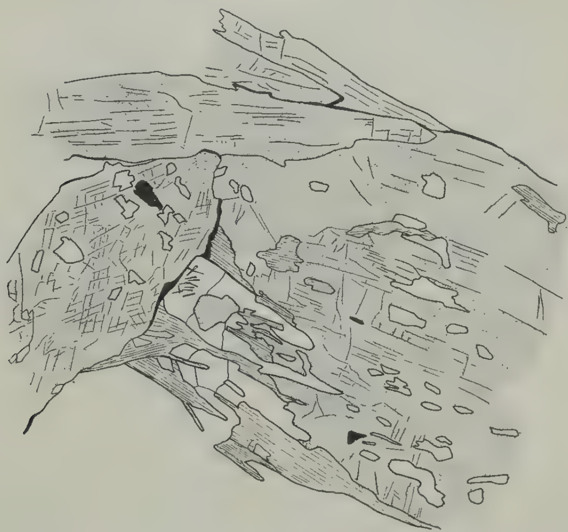


Fig. 8. Illustrating the replacement of actinolite by quartz and biotite. Note how replacement is controlled by the prominent (110) cleavages of the amphibole (× 400). Actinolite (stippled), quartz (white), biotite (ruled).

Biotite, with strongly pleochroic haloes about enclosed zircons, is always abundant and occurs in close association with schorlite which it replaces. Quartz and plagioclase felspar compose the light-coloured foliae. Acid plagioclase (An_9Ab_{91}) constitutes the principal felspar, and is occasionally prominently developed in the mineralised shear zones. Subsidiary andesine ($An_{40}Ab_{60}$), however, is confined to a zone comprising part of the highly contorted biotite-rich schist outlined in the cross-sections illustrated in Figs. 3A and 3B. A notable feature of the more basic felspar is the frequency of acline twins. This substantiates the view of F. J. Turner (1951) that "pericline twinning of plagioclase is at least as common as albite twinning in rocks of higher metamorphic grade". A determination of a typical acline twin by means of the U. M. stage yielded the following optical constants:

$2V\gamma = 85°$ composition plane (001), twin axis [010] cleavage (010), composition $An_{38}Ab_{62}$.

Unreplaced amphibole vestiges consist of anthophyllite, actinolite and tremolite. Anthophyllite, however, frequently substitutes for the other amphiboles in these highly sheared rocks. The fibrous amphibole remnants are distinctly orientated parallel to the prominent S-planes (see Fig. 9).

Almandine garnet, occurring as idioblasts with an average diameter of 1 mm. are occasionally present in the schist. These garnetiferous bands are usually confined to discreet horizons and are rarely extensive.

Fig. 9. No. 2 Shear Vein ($\times$ 375). Note the prominent biotite-tourmaline and quartz rich foliae. Ore grains and anthophyllite are distinctly orientated parallel to the dominant S-plane structure. Anthophyllite (stippled), biotite ruled), tourmaline (dotted), quartz (white), ore (black).

Aplite. A fine-grained equigranular rock with a granitoid texture. In thin section the rock is seen to be composed of quartz and albite-oligoclase felspar. The oligoclase is characterised by occasional extremely fine twin lamellae and a $2V\gamma = 88°$. Clusters of alkali tourmaline, pleochroic from pale blue to green are usually present. The tourmaline (elbaite) occurs as skeletal crystals, frequently in optical continuty, replacing quartz and felspar. Muscovite and calcite are the main accessory constituents.

Dolerite. A well defined chill zone $\pm$ 4.0 feet wide is composed of prominent plagioclase phenocrysts ($An_{58}Ab_{42}$) set in a microcrystalline groundmass of pyroxene, plagioclase ($An_{44}Ab_{56}$) and iron ore.

The dyke is a medium-grained rock. Sub-ophtic augite with $2V\gamma = 40°$, $\gamma/c = 39°$ and $\gamma = 1.721$ ($\pm$ 0.003) is the dominant pyroxene. Two generations of plagioclase felspar were noted. The early plagioclase exhibits continuous zoning from core to margin, ranging in composition from $An_{70}Ab_{30}$ to $An_{55}Ab_{45}$ respectively. The unzoned second-generation plagioclase, which is more abundant, has an average composition of $An_{40}Ab_{60}$. This rather low

PERIDOTITE
(Jamestown Igneous Complex)

REGIONAL METAMORPHISM

Actinolite - Tremolite Amphibolite

CONTACT METAMORPHISM	**Contact Amphibolite**	**Weakly Sheared Rock**	**Strongly Sheared Rock**	
Early Pneumatolytic Metasomatism	Minor schorlite, apatite and scheelite replacing amphibole	Moderate schorlite, apatite and trace of scheelite replacing amphiboles	Abundant schorlite, minor apatite	
High temperature Hydrothermal Alkali-Si-Metasomatism	Phlogopite replacing amphibole → **Altered Amphibolite**	Minor biotite and quartz and some albite - oligoclase felspar replacing amphibole → **Amphibole schist**	Abundant biotite introduced. Some quartz and albite-oligoclase felspar. Extensive replacement of amphiboles → **Quartz - tourmaline - biotite - felspar schist**	Aplite veins appear along the continuation or projected continuation of shear zones. Quartz and albite-oligoclase felspar
Low temperature Hydrothermal. CO_2, H_2O, Metasomatism	Minor development of chlorite and talc	Minor development of talc	Development of talc-carbonate-chlorite schists in shear zones unaffected by earlier metasomatic agencies	

Fig. 10—Schematic representation of postulated metamorphic processes.

anorthite content is regarded by J. W. Brandt (1948) as characteristic of the dolerites of the N. E. Transvaal. The felspars are twinned according to the Albite, Roc Tourné and Ala laws. Accessory iron ore is mantled by biotite and chlorite. A mesostasis of micropegmatite represents the last phase of crystallisation.

B. Petrogenesis.

In the previous section the mineralogical and textural variations relating to the transformation of amphibole into a varied assortment of amphibole schists have been discussed. The probable genesis of such rock types from an original ultrabasic intrusive mass by the agencies of regional and contact metamorphism are outlined in Fig. 11 and may be considered as follows:

Regional Metamorphism.

The Klein Letaba amphibolite occurrence displays a remarkably limited mineral composition, characterised by actinolite and tremolite, implying a derivation from a specific rock type. H. H. Hess (1933) writes: "The extreme ultrabasics, dunites, peridotites and pyroxenites which contain no plagioclase, have a very low alumina content. As a result of this the mineral facies developed are much more limited than in those rocks which contain a little plagioclase. With no alumina present, the aluminous amphibole, hornblende, cannot form and thus hornblende amphibole facies will be lacking. The actinolite amphibolite facies will develop directly from the original minerals of the ultrabasic."

Field and laboratory evidence supporting the view that the actinolite amphibolite at Klein Letaba was derived from the regional metamorphism of an ultrabasic intrusive may be summed up as follows:

(a) The lenticular masses of serpentine with associated magnesite situated to the south of the mine are considered by Van Zyl (1942) to be derived from an original ultrabasic rock type rich in olivine, such as dunite or harzburgite. The transformation of the dunite into serpentine and magnesite may be illustrated by the following chemical equations:

$$\underset{\text{Olivine}}{3Mg_2Fe_2Si_2O_8} + 4H_2O + O_2 = \underset{\text{Serpentine}}{2H_4Mg_3Si_2O_9} + 2Fe_3O_4 + 2SiO_2$$

$$\underset{\text{Serpentine}}{H_4Mg_3Si_2O_9} + 3CO_2 = \underset{\text{Magnesite}}{3MgCO_3} + 2SiO_2 + 2H_2O$$

(b) An analysis of relatively unaltered amphibolite by means of the spectrophotometer yielded a value of 2.60% Al_2O_3. A micrometric analysis of this sample indicates a mineral composition corresponding to 89% actinolite and 11% penninite. Though the amphibolite is somewhat altered the low Al_2O_3 content tends to substantiate the field association pointing to a derivation from an ultrabasic source, and lends support to the views expressed by H. H. Hess quoted above.

(c) No possible stratigraphic linkage of the amphibolite to any possible parental sedimentary horizon was disclosed by surface mapping.

Towards the close of the period of regional metamorphism, and before

the consolidation of the neighbouring granite gneiss, intense dynamic deformation of the amphibolite resulted in the development of well defined shear zones. These shear zones acted as channelways for diverse metasomatic agencies, which are mainly responsible for the alteration of the amphibolite. The metasomatism is regarded as a phase of contact metamorphic action induced by the flanking granite gneiss which is intrusive. The possibility of the existence of an underlying later batholithic intrusion which has not yet been recognised or exposed in this inadequately investigated region should, however, be borne in mind.

Mention may here be made of the breakdown of the actinolite and tremolite structures under the influence of dynamic stress. The formation of calcite and an aluminium-poor anthophyllite, as a consequence of this deformation, preceded the main influx of metasomatic agents. The development of limited amounts of andesine felspar, noted in discreet horizons within the contorted biotite-quartz-amphibole schist, must also be regarded as the result of regional metamorphism. The frequency of acline twinning in these felspars would tend to substantiate this view.

Contact Metamorphism.

Intense metasomatic replacement appears to have been the principal contact metamorphic process that affected the rocks comprising the mine suite. The nature of the replacing vapours and fluids is in accord with a derivation from a granite source, and the loci of replacement are subject to a marked structural control. A study of the metasomatic products reveals a transition from pneumatolytic to high and low temperature hydrothermal conditions of replacement.

The term metasomatism as used in this discussion is based on the definition by V. M. Goldschmidt (1922) that "Metasomatism is a process of alteration which involves enrichment of the rock by new substances brought in from the outside. Such enrichment takes place by definite chemical reactions between the original minerals and the enriching substances."

The following classification of the metasomatic processes arranged from early pneumatolytic to late hydrothermal is supported by petrological evidence:

(a) *Pneumatolytic Metasomatism.* Though some measure of gas or vapour action is noted throughout the metasomatic process, the early pneumatolytic stage is responsible for the largest concentration of the volatile constituents. The introduction and fixing of B, F, P, W, Ti as complex ions with an acid reaction as envisaged by N. Bowen (1933) resulted in the reactive replacement of amphibole by tourmaline (schorlite), apatite and scheelite. The formation of apatite and scheelite does not necessarily imply introduction of carbonate at this stage, since adequate quantities of combined and released lime were available in the fresh and mechanically deformed magnesian and lime amphiboles of the invaded country rock. The presence of tourmaline and apatite in the amphibolite wall rock and amphibole schist with poorly developed S-plane structures, at some considerable distance from the major shear zones, is believed to favour the hypothesis of early gas-phase action rather than the activity of an initial alkaline liquid medium as suggested by F. G. Smith (1946).

The large scale tourmalinisation noted within the shear zones is to be expected, as these provided the main channelways for the release of pressure and influx of the pneumatolytic and subsequent metasomatic agents.

The absence of typical pegmatite development in association with the pneumatolytic phase may be accounted for by the following points of view:

P. Niggli (1929) in his dissertation on the physical chemistry of ore-generating magmas points out that with the escape of the volatile constituents from a melt crystallising in an open system (i.e. escape of volatiles may be facilitated by fractures in the roof of the magma chamber), the pegmatitic stage of crystallisation may be eliminated.

F. G. Smith (1948) provides experimental evidence which tends to support the view that "near the end of magmatic crystallisation there is the generation of a water solution which is immiscible with the residual silicate solution, and soluble components are partitioned between the two liquids. The crystallisation of the silicate liquid in channels of escape from the magma forms pegmatite veins, and the crystallisation of the water solution in similar channels forms veins containing the metallic sulphides." The author further indicates that the ratio of the viscosities of the two liquids is approximately 10^5:1 and 10^8:1, and deduces that the hydrothermal solution may migrate 10 kilometers up a fissure in the time taken for the pegmatitic solution to move but a few centimeters.

The above opinions tend to substantiate each other, and imply a suppression of the pegmatitic phase under specific P-T conditions.

(b) *Alkali and Silica Metasomatism.* The high temperature hydrothermal solutions, which invaded the shear zones towards the close of the pneumatolytic phase, are considered to have acquired a distinct alkaline character after the escape of the acid radicals in the form of gases and vapours. The main influx of the metallic sulphides is also thought to have taken place during this hypothermal stage, but the discussion of this aspect of the problem may conveniently be deferred to a later section.

Introduction into the shear structure of K, Na, Al and some Si as simple ions carried in alkaline solution, induces large-scale metasomatic replacement of the amphiboles with development of biotite, quartz and acid plagioclase felspar. It is noteworthy that, in contrast to the agents of pneumatolytic metasomatism, the alkali-silica metasomatic phase is much more restricted to the confines of the shear structures, though some phlogopite does replace the amphibole in the adjoining wall rock.

Vein-like bodies of aplitic composition located on the continuation or projected continuation of the metasomatised shear-zones may tentatively be regarded in the light of residual products of hypothermal alkaline solution depleted in volatiles. The characteristic progressive enrichment in soda of the later and lower temperature pegmatitic-hydrothermal magmatic residues is here displayed by the exclusive appearance of green soda-rich tourmaline (elbaite) in the aplitic bodies as opposed to the iron-rich tourmalines (schorlite) which characterise the early pneumatolytic phase.

The following examples from the literature are considered by the writer to fully substantiate the view that large scale metasomatism of dynamically deformed amphibolite gives rise to a varied mineral assemblage of biotite-rich schists:

(i) A. F. Buddington (1939) with regard to the metamorphic geology of the Adirondacks states: "The arteric migmatites of amphibolite and granite

pegmatite are found to vary from banded arterites in which there has been mechanical injection of granitic magma along the foliation planes, through facies in which the amphibolite has been completely changed to a biotite schist and the veinings are due to replacement by pegmatitic solutions, to uniform rocks that have completely altered and recrystallised by permeating solutions and materially changed in composition. In the latter two types, the common formation of scapolite and the abundance of minerals such as apatite, sphene, magnetite, tourmaline and some zircon suggests that fluids highly charged with mineralisers were responsible for the transformations."

(ii) With reference to the metasomatism associated with the greenstone-hornfelses of Kenidjack and Botallack, Cornwall, C. E. Tilley (1935) states: "It is considered that hot solutions eventually emanating from a granitic source have passed through the greenstones and produced an effective removal of lime with the production of anthophyllite and related types of hornfels. The process is in reality one of contact metasomatism. It is believed that while the chief substance removed has been lime, the process has not been one of simple abstraction, but of replacement in which the constituents removed have been replaced by other material. As already noted there has been clear introduction of potash resulting in the formation of much biotite in some of the rocks."

(iii) Th. A. Dodge (1942) makes the following comment with relation to the altered amphibolites of the lead area, Northern Black Hills, South Dakota: " . . . the presence of hornblende and biotite in random orientation in both the marginal schists and the amphibolite point to high-temperature hydrothermal action after (but probably not long after) the cessation of all folding movement."

In conclusion the writer wishes to point out that the pronounced difference in the chemical and mineralogical composition between the amphibolites and the biotite-quartz-amphibole schists rules out the process of metamorphic differentiation as having been effective in the development of these schists. By the term metamorphic differentiation the author implies the formation of pseudo-stratification in rocks without introduction of material, as applied by F. J. Turner (1941) to the developments of schists at Otago, New Zealand.

(c) *Low Temperature Hydrothermal Action.* The writer believes that the talc-carbonate-chlorite and antigorite schists were formed as the products of water and carbon dioxide metasomatism under low temperature hydrothermal conditions. This view accords with that expressed by F. E. Keep (1929), that the Rhodesian talcose rocks owe their origin to the contact action of an adjacent granite. H. H. Hess (1933) believes that the steatite deposits at Schuyler, Virginia, were produced by similar low temperature hydrothermal solutions acting upon actinolite-rich amphibolites.

VI. MINERAGRAPHY OF THE ORES

Among the minerals recognised under the reflecting microscope were ilmenite, lollingite, arsenopyrite, pyrrhotite, pentlandite, pyrite, chalcopyrite, sphalerite, gold and electrum.

Ilmenite. This mineral is irregularly distributed in accessory amount in both the pyrrhotite-rich and arsenopyrite-rich ores. Ilmenite generally occurs as irregular crystals that are interstitial to the silicates. Occasional

development of crystal grains which exhibit a distinct rod-like habit were noted. These rods attain an average size of 0.10 x 0.01 mm., indicating a length to breadth ratio of approximately 10:1. Excellent polish, low reflectivity and an anisotropism under crossed nicols from grey to brown in addition to weak reflection pleochroism characterise this mineral.

Ilmenite, which was noted in contact with arsenopyrite and sphalerite, is clearly replaced by the latter minerals, and is considered to be the earliest mineral in the paragenesis of the ores.

Lollingite. This mineral occurs as skeletal crystals which are extensively replaced by arsenopyrite. In the arsenical ores, lollingite may constitute up to 5% by volume of the ore constituents. The rod-shaped crystals attain an average grain size of 0.13 x 0.03 mm. The observation that the skeletal crystals of lollingite are often in optical continuity within the replacing arsenopyrite, indicates that they comprised larger and more extensive crystals before the introduction of the arsenopyrite. Identification of lollingite was rendered difficult because of its constantly intimate association with arsenopyrite, with which it has many physical properties in common. Positive identification of lollingite, however, was effected by the following experimental applications:

(a) *Chromographic contact-print method.*

D. Williams and F. M. Nakhla (1951) present a detailed account of the technique involved. The basic principle is the same as that underlying spot testing, but whereas in spot testing or the microchemical methods the presence of an element is established only at a particular point on the surface, the chromographic method gives a permanent record of the spatial distribution of a particular element over the entire surface of the polished section. The method may briefly be described as follows.

Gelatine coated paper is prepared from ordinary photographic glossy bromide paper by soaking the latter in a standard hypo-solution for approximately 15 minutes. This operation is to be carried out in a safe light. After thorough washing and drying the paper is ready and may be used in two ways.

(i) *Simple contact-print method.* The gelatine coated paper, cut to a size slightly larger than the specimen to be tested, is soaked for several minutes in an appropriate attacking reagent (either acids or strong bases). The gelatine side of the saturated paper is placed against the polished surface of the ore mount, good contact being assured by applying moderate pressure (a standard Leitz press was used for the purpose). The length of attack varies from a few seconds to several minutes depending upon the minerals involved. The paper is then stripped from the surface of the ore mount and placed immediately in an appropriate reagent. A colour pattern develops which shows the distribution of the element in question throughout the entire contact area of the polished section.

The intimately associated arsenopyrite and lollingite relationship was investigated by testing for the presence and distribution of arsenic and nickel as follows.

Test for arsenic.

Attacking reagent:	$NH_4OH + H_2O_2$ (1/1)
Specific reagent:	$AgNO_3$ (1%-2% solution)
Colour of print:	Brown

Test for nickel.

Attacking reagent: NH_4OH + rochelle salt 10% + dimethylglyoxime

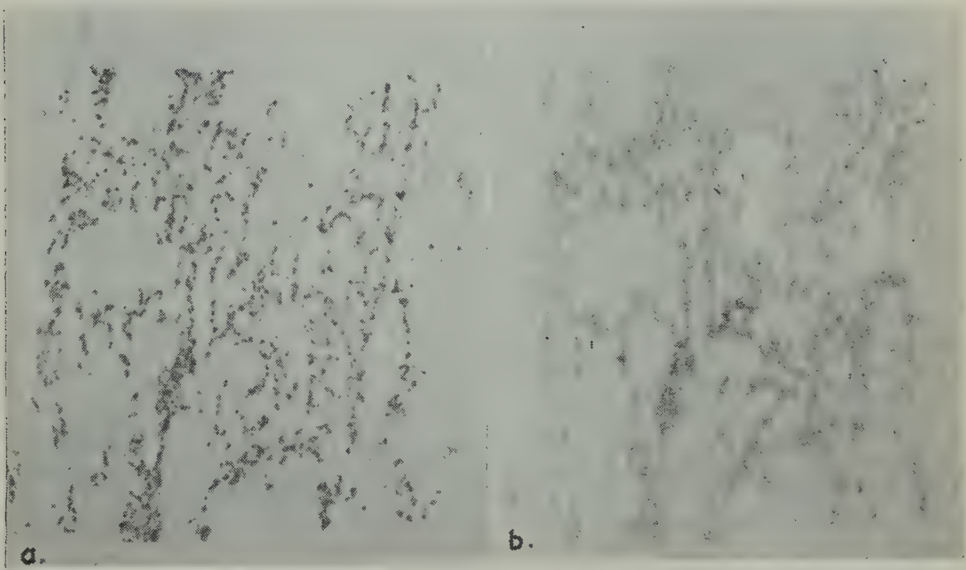
No specific reagent required — direct print with red colour.

After prolonged experimentation during which he varied the time of contact to over 15 minutes, the writer had no success with the application of the simple contact-print method.

(ii) *Electrographic contact-print.* The procedure follows the simple contact-print method described above, but a current of 6 volts applied for 60 seconds was employed in the attacking operation. Contact with the polished specimen was obtained by drilling a hole in the base of the bakelite mould so that electrical contact with the undersides of the ore specimen could be made. This electrical connection is fixed to the positive pole of the battery. A metallic foil wrapped around a glass slide is placed on top of the gelatine coated paper and is connected to the negative pole of the battery. The writer found that the smoother the surface of contact between the metallic foil and the paper, the sharper the resultant colour print, even with moderately applied manual pressure on the plunger of the press. When this technique was used, exceedingly sharp well defined contact prints were obtained.

The accompanying photograph (Ph. 1) shows the arsenic and nickel distribution of the arsenopyrite-lollingite ores, and the correspondence of detail with regard to the distribution of these elements is strikingly apparent.

Care must be taken in attempting to make quantitative estimates by using this method, as the depth of colour is proportional to the amount of material dissolved rather than to the quantity of the element present.



Ph. 1. Chromographic prints. (a) Arsenic pattern; (b) Nickel pattern.

(b) *X-ray diffraction method.*

By means of a standard dental drill, a small quantity of the intimately associated arsenopyrite-lollingite ore was obtained from a polished ore specimen.

The powder was mounted in a 57.4 mm. diameter X-ray powder camera. The X-ray powder pattern was prepared with the use of Cu-Kα radiation Similarly, X-ray diffraction powder patterns were obtained from standard arsenopyrite and lollingite specimens. The comparative powder patterns are listed in Table I. The X-ray diffraction pattern of the arsenopyrite-lollingite ore from Klein Letaba contains lines appearing in both the standard arsenopyrite and lollingite patterns, suggesting that the ore under investigation contains a mixture of these two minerals.

Arsenopyrite-lollingite ore (Klein Letaba)		Arsenopyrite (Standard)		Lollingite (Standard)	
d	Intensity	d	Intensity	d	Intensity
2·81	1				
2·60	4	2·60	5	2·60	10
				2·58	10
2·40	10	2·40	10		
				2·35	6
				2·30	
2·00	4	1·90	1	1·90	1
1·82	7	1·81	7	1·84	10
		1·75	1		
				1·67	8
1·63	2	1·62	4	1·63	8
1·54	1	1·54	4	1·57	1
				1·49	2
1·45	1			1·45	2
1·40	1	1·40	1		
1·35	1	1·34	1		
		1·27	1	1·28	1
		1·22	1		
1·15	1	1·17	1	1·15	1
				1·14	1
1·10	1			1·10	1
1·05	1			1·05	2
1·00	1	1·00	1		

TABLE I

Comparison of Klein Letaba arsenopyrite-lollingite X-ray diffraction patterns with the patterns obtained from standard arsenopyrite and lollingite

(c) *Qualitative Spectrochemical Analysis.*

A qualitative spectrochemical analysis of the ore rich in arsenopyrite-lollingite yielded characteristic spectral lines for nickel and cobalt.

The above experimental evidence would indicate a lollingite containing traces of nickel and cobalt, tentatively placing the mineral in the nickelian lollingite field as envisaged by R. J. Holmes (1947).

Further optical properties revealed by examination under the reflecting ore microscope may be listed as follows:

Pleochroism Weak (in oil): whitish to yellowish white

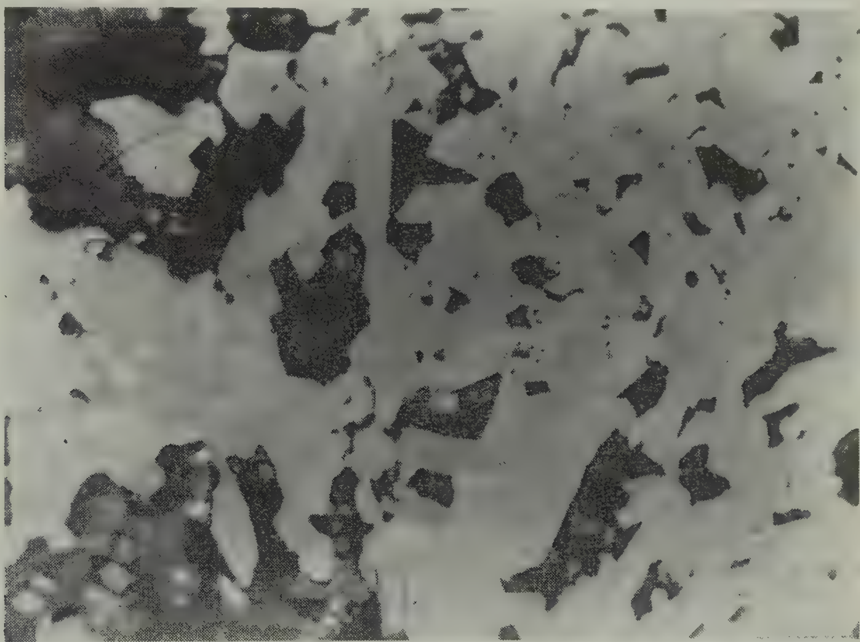
Anisotropism marked (in oil)

+ Nicols: greenish grey to salmon red

+ Nicols (-5°): greyish blue to red brown.

A reflectivity determination on the intimately associated arsenopyrite-lollingite ore (containing roughly equal proportions by volume of arsenopyrite and lollingite) which covered the entire microscope field yielded a reflectivity measurement of 54% using a photometer ocular and green filter (in air). This value is intermediate in reflectivity when compared with the values 52.4% and 57% listed respectively for arsenopyrite and lollingite by Uytenbogaardt (1951). Corresponding reflectivity values for arsenopyrite and lollingite quoted by P. Ramdohr (1950) are 51.0% (incorrectly labelled in text!) and 57% respectively.

Arsenopyrite. This mineral is the dominant constituent in the arsenic-rich ores. Apart from occurring as irregular masses intimately replacing lollingite, arsenopyrite is developed as euhedral crystal aggregates which are frequently oblong or diamond shaped in cross-section. The extensive replacement of lollingite by arsenopyrite is illustrated in photomicrograph (Pm. 1). The mineral is distinguished from lollingite by its feebler reflection pleochroism, less intense anisotropism and slight difference in hardness.



Pm. 1. Arsenopyrite (pale grey) — Lollingite (white) with Silicates (black), Crossed Nicols, ×100.

Both arsenopyrite and lollingite are unaffected by etching with FeCl (M. N. Short, 1940). The writer found, however, that the standard Mericas solution may usefully be employed to distinguish between lollingite and arsenopyrite, as the former mineral is unaffected, whereas the latter etches to a green-yellow iridescence which does not rub off easily.

The replacement of lollingite by arsenopyrite, which is so clearly displayed in the arsenical ores, points to the early crystallisation of the di-arsenide lollingite, followed by arsenopyrite. This conflicts with the view expressed by Schneiderhohn and Ramdohr (1931), who state that lollingite is almost always formed later than arsenopyrite, and refer to an occurrence at Reichenstein, where arsenopyrite is younger, as exceptional.

But evidence supporting early generation of lollingite is given by F. L. Stillwell and A. B. Edwards (1939), who point out that in the Broken Hill lode, lollingite is earlier than arsenopyrite and is replaced by the latter. These authors regard this as the normal sequence and state that similar relationships between lollingite and arsenopyrite were observed in the Bon Accord Mine, the Gorden Gold Mine, and the Sulphide Junction Mine (New South Wales). They conclude that the evidence "implies that originally arsenical ore solutions gave place to sulphide ore solutions in the course of the development of the lodes." More recently J. S. Stevenson (1951) has shown that in the Little Gem and Victoria uraninite deposits, lollingite crystallised after uraninite, and is extensively replaced by arsenopyrite.

Recent observations, therefore, not only substantiate the writer's views with regard to the early crystallisation of lollingite in the Klein Letaba ores, but tends to cast some doubt on the view expressed by Schneiderhohn and Ramdohr.

To the writer's knowledge this is the first occurrence described from the Union of South Africa where lollingite is directly associated with a gold ore. But mention should be made of the occurrence of lollingite in the mineralised zones associated with the Stellenbosch-Kuils River granite pluton described by D. L. Scholtz (1946).

Pyrrhotite. This mineral constitutes the principal sulphide of the disseminated ores, and generally predominates over arsenopyrite in the more massive ore lenses. The crystal outlines are usually irregular and interlocking. Euhedral crystals are rarely observed.

Pyrrhotite extensively envelopes arsenopyrite (Fig. 11). The observation of a large optically continuous crystal of pyrrhotite enclosing small scattered crystals of arsenopyrite is regarded as convincing evidence for the later crystallisation of pyrrhotite. Pyrrhotite also encloses pyrite, which it is believed to replace.

The optical properties revealed under the ore microscope may be summarised as follows:

Pleochroism: Not very distinct — creamy to orange cream.

Anisotropism: Strong.

+ Nicols (air): Yellowish grey to bluish grey.

Reflectivity: Photometer ocular, green filter (in air) = 37%.

Incipient alteration of pyrrhotite along small fractures is typical of the massive ore lenses. The alteration develops both along, and at right angles to, the fractures, and may assume various orientations within the same pyrrhotite crystal. This alteration is regarded as the result of late hydrothermal action. The view that these fractures are the result of crystal growth in a confined space is borne out by the fact that fractures are entirely lacking in the disseminated pyrrhotite occurring in the mineralised shear zones.

These areas of alteration should not be confused with the "flames" of exsolved pentlandite that characterise the pyrrhotite of the disseminated

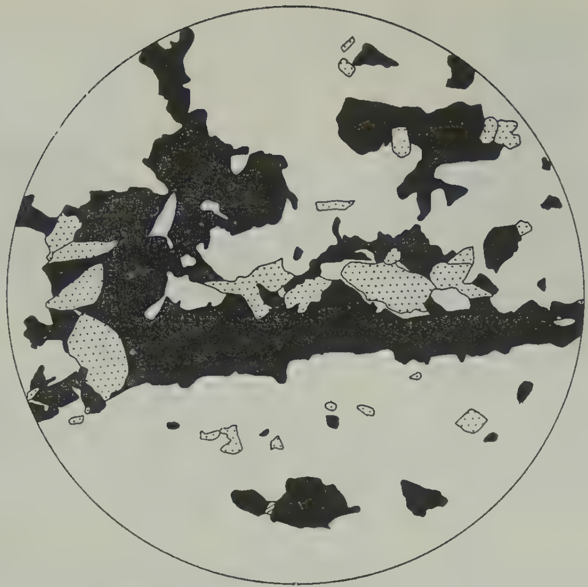


Fig. 11. Camera Lucida drawing of polished section of pyrrhotite-rich ore. × 200. Pyrrhotite (black), arsenopyrite (dotted), chalcopyrite (ruled). Gangue (white).

ore, which are never observed in the more massive pyrrhotite-rich ore lenses. The author found that etching with potassium dichromate served to distinguish these alteration zones from pentlandite, as the former tarnishes readily whereas the latter is unaffected.

Disseminated pyrrhotite occurring in the highly contorted biotite-quartz-amphibole schist is entirely free from any strain foliae, which indicates that crystallisation of sulphide occurred after cessation of the major folding movements. A pronounced pseudomorphic replacement texture is in evidence in the mineralised shear zones, as the pyrrhotite crystals are distinctly orientated parallel to the prominent S-plane structure (see Fig. 9).

Pentlandite. Exsolution secondary "flames" or "brushes" of pentlandite are restricted to the pyrrhotite of the disseminated ores, and are developed along definite crystallographic directions of the pyrrhotite host. Chalcopyrite occasionally replaces pyrrhotite along a pyrrhotite-pentlandite "flame" contact (Fig 12). It is noteworthy that no primary pentlandite is present in the ores.

The association of "secondary" pentlandite "flames", specifically with disseminated pyrrhotite ores, and the complete absence of these "flames" in the more massive pyrrhotite ores are striking features which throw some light on the possible paragenesis of the pentlandite "flames" in the hypothermal ores of the Klein Letaba type.

The view expressed by Newhouse (1927) that narrow blade-like masses of pentlandite within pyrrhotite crystals may be formed by slow cooling of a pyrrhotite-pentlandite solid solution, is hardly applicable to the relatively narrow ore zones of the Klein Letaba type.

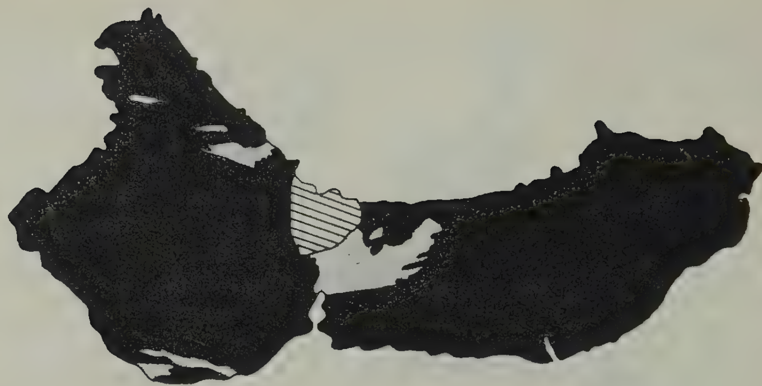


Fig. 12. Camera Lucida drawing of polished section of disseminated ore. × 338. "Flames" of secondary pentlandite developed in two large grains of pyrrhotite. Pyrrhotite (black), pentlandite (stippled), chalcopyrite (ruled), Gangue (white).

More recently, however, an investigation of the Fe-Ni-S system by J. E. Hawley, G. L. Colgrove and H. F. Zurbrigg (1943) has materially advanced our knowledge about the possible paragenesis of pyrrhotite and pentlandite. These authors could find no experimental evidence supporting the view of partial or complete immiscibility in the case of a melt of composition $(NiS)_4(FeS)_6$.

Their investigation further revealed that the solubility of NiS in FeS is considerably influenced by the amount of sulphur present in the system, the sulphur tending to increase the solubility of NiS in FeS. A nickeliferous pyrrhotite with a wide range in nickel content is therefore postulated by these authors. Recently, V. A. Haw (1948) has shown that loss of sulphur from such nickeliferous, artificially prepared pyrrhotites, by heating in sealed tubes, resulted in the formation of "flame-like" segregations of pentlandite.

The presence of cobalt and nickel in natural pyrrhotites is now well established. K. Rankama and Th. G. Sahama (1949) quote that the average nickel content of 92 pyrrhotite samples from the Finnish pre-Cambrian rocks is 0.20%.

In order to prove the presence of nickel in the massive pyrrhotite ore at Klein Letaba (i.e the pentlandite "flame-free" ore), the writer made a qualitative spectrochemical analysis, which revealed distinct nickel lines of the following wave-lengths:

2821.29A°, 3003.63A°, 3120.004A°, 3050.819A°, 3054.326A°, 3461.65A°, 3458.47A°.

It is significant that the massive pyrrhotite ore (which may constitute 36% by volume of the ore zone) may contain appreciable quantities of pyrite, whereas the disseminated pyrrhotite ore (constituting up to 2.2% by volume of the ore zone) is invariably "pyrite free". It is generally accepted that development of pyrite presupposes a high concentration of sulphur, possibly H_2S. (H. Vayrynen, 1955).

The writer considers, therefore, that the massive pyrrhotite ore, owing to conditions favourable to the chemical activity of sulphur, was able to hold substantial amounts of nickel in solid solution. The disseminated pyrrhotite, however, occurring within the more open shear structures, could not retain sufficient sulphur to render the NiS soluble, with resultant shedding of nickel sulphide as exsolution "flames" of secondary pentlandite.

Pyrite. Scattered irregular crystals and crystal aggregates of pyrite are associated with the massive pyrrhotite ores. The crystals are frequently enveloped by pyrrhotite, suggesting that pyrite precedes pyrrhotite in the paragenetic sequence.

Occasionally the pyrite exhibits a weak anisotropism between crossed nicols. This somewhat anomalous anisotropism according to W. Uytenbogaardt (1951) may be ascribed to internal strain and tension caused by some iron sulphide surplus or arsenic admixture.

Sphalerite. This mineral, present in accessory amount, is best developed in the high-grade gold ores. Characterised by low reflectivity and intense red to reddish-brown internal reflection, sphalerite frequently envelops small grains of pyrrhotite, which it also replaces along intergrain pyrrhotite boundaries. Blende also replaces arsenopyrite, suggested by the scalloped appearance of the arsenopyrite host, which appears to have been "bitten-into" by the sphalerite.

The sphalerite crystals are occasionally twinned, and may exhibit a weak anisotropism. The anisotropism of certain sphalerite specimens is referred to by P. Ramdohr (1950), who states that "Einige sehr eisenreiche Proben waren einwandfrei deutlich anisotrop."

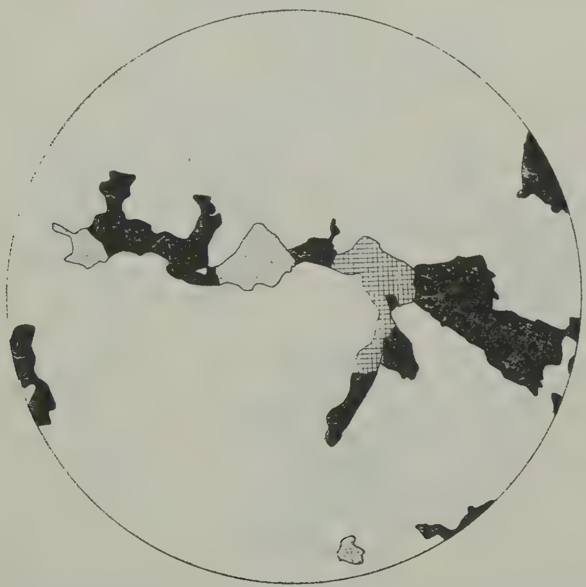


Fig. 13. Camera Lucida drawing of a polished section illustrating the replacement of pyrrhotite by gold and sphalerite. × 200. Pyrrhotite (black), gold (stippled), sphalerite (cross hatched), gangue (white).

Chalcopyrite. This mineral frequently accompanies gold in the high-grade ores. Apparently sphalerite, chalcopyrite and gold were late in the paragenetic sequence. Chalcopyrite was noted in the following mineral associations and relations:

(a) In contact with pyrrhotite, which it clearly replaces along well defined fractures within the pyrrhotite host.

(b) In contact with and replacing pyrrhotite along pyrrhotite-silicate intergrain boundaries.

(c) In contact with and replacing pyrrhotite and secondary pentlandite along intergrain boundaries of the latter.

(d) In contact with and replacing arsenopyrite along intergrain boundaries.

Gold. This mineral occurs as grains varying in size from approximately 0.005 mm. to a maximum of 0.23 mm. It is considered to belong to a late stage in the paragenetic sequence, and was noted in the following associations.

(1) Free as tiny veins and grains replacing the silicate gangue.

(2) Replacing sphalerite and chalcopyrite.

(3) Replacing pyrrhotite, arsenopyrite and lollingite along intergrain boundaries.

Figs. 13 and 14 illustrate some typical gold-sulphide and gold-silicate relationships.

Fig. 14. Camera Lucida drawing of a polished section showing pyrrhotite, arsenopyrite, chalcopyrite and gold relationships in high grade ore. × 200. Pyrrhotite (black), arsenopyrite (dotted), chalcopyrite (ruled), gold (stippled), gangue (white).

Electrum. A few minute grains of a creamy-white mineral exhibiting an exceedingly high reflectivity were noted in contact with sphalerite and pyrrhotite in one ore specimen. Though they are too small to permit of positive identification, the writer tentatively regards these grains as electrum.

VII. GENERAL CONSIDERATIONS REGARDING THE DISTRIBUTION, PARAGENESIS AND GENETIC CLASSIFICATION OF THE KLEIN LETABA ORES

A. Ore Types and Distribution.

The gold-bearing horizons at the Klein Letaba Mine may readily be divided into two main types, namely the mineralised quartz-amphibole schists (which characterise the major shear zones) and the massive quartz-sulphide lenses which are generally confined to, or lie close to and parallel to, the main shear structures. The former ore type constitutes the typical disseminated variety in which pyrrhotite is invariably the dominant sulphide. In the more massive sulphide lenses, on the other hand, pyrrhotite or asenopyrite constitute the dominant hypogene ore constituents. The relative percentage volume distribution of the principal hypogene sulphide minerals comprising the No. 2, 3 and 4 ore zones are outlined in Table II.

Although the more massive sulphide lenses constitute a relatively small proportion of the potential mineable tonnage, they are of considerable economic significance in that the high-grade gold shoots are restricted to these horizons.

VEIN-Type	MAIN CONSTITEUNTS			ACCESSORY CONSTITUENTS
	% Volume Pyrrhotite	% Volume Arsenopyrite-Lollingite	% Volume Gangue	% Volume ilmenite and/or chalcopyrite and sphalerite
No. 2 Vein, Disseminated ore ..	2·20	0·30	97·50	Traces
No. 2 Vein, Pyrrhotite-rich ore ..	36·10	Nil	63·90	Traces
No. 2 Vein, Pyrrhotite-rich ore ..	28·60	1·20	70·20	Traces
No. 2 Vein, Pyrrhotite-arsenopyrite ore	27·60	11·80	59·90	0·70
No. 3 Vein, Pyrrhotite-rich ore ..	21·10	Nil	78·60	0·30
No. 3 Vein, Arsenopyrite-rich ..	1·10	5·60	93·30	Traces
No. 4 Vein, Pyrrhotite-rich	8·10	1·20	90·10	0·20
No. 4 Vein, Arsenopyrite-rich ..	18·60	38·20	43·20	Traces

TABLE II
Micrometric analyses of the ores constituting the Nos. 2, 3 and 4 Veins

There appears to be a marked relationship between the tenor of the gold ores and the vein width. This feature is particularly apparent in the cross-section through boreholes KL11 and KL22 (see Fig. 3B), where it is noted that massive ore lenses are restricted to the tapering apex of the No. 4 Vein and the narrow No. 3 Vein, in contrast to the relative absence of any sulphides in the wider portions of the No. 2 and No. 4 Veins between the 400-foot and 700-foot levels.

B. **Paragenesis.**

The Klein Letaba ore deposits are considered to be genetically related to and contemporaneous with the phase of alkali-silica metasomatism. The marked structural control, induced by strong shearing, which has already been stressed with regard to the metasomatic action, applies equally forcefully to the localisation of the ore constituents.

The succession of ore minerals from earliest to latest as determined in a study of polished sections of the ores is as follows: ilmenite, lollingite, arsenopyrite, pyrite, pyrrhotite, sphalerite, chalcopyrite, gold and electrum. Pentlandite bears an exsolution relationship to pyrrhotite. Thin section studies

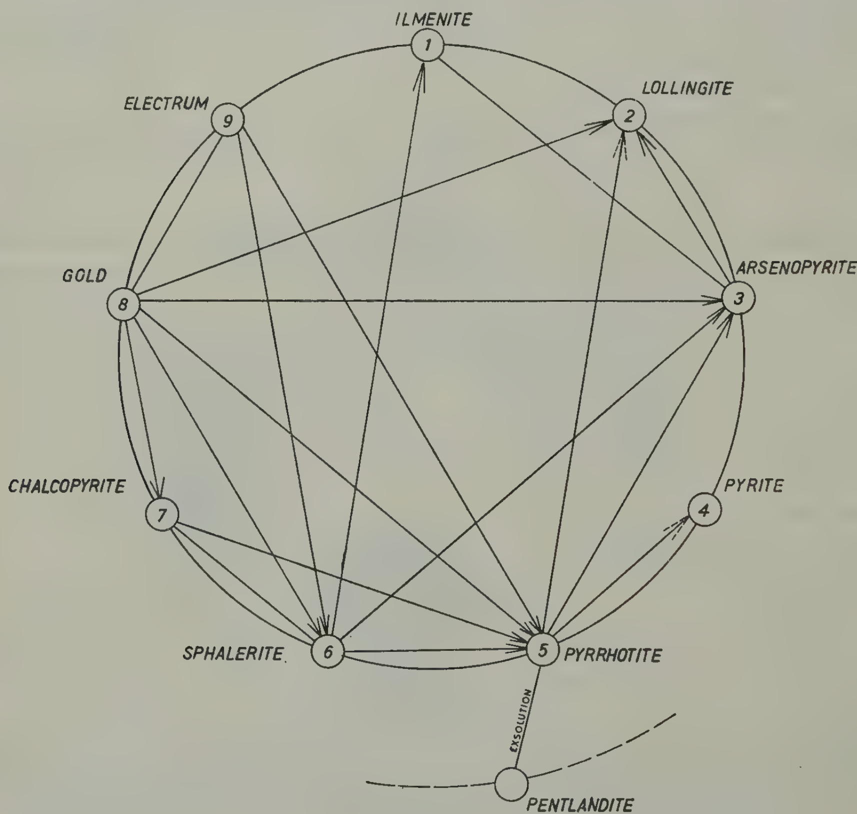


Fig. 15.

of the mineralised zones indicate that the ore minerals crystallised later than the gangue silicates which they clearly replace.

The paragenetic relations of the Klein Letaba ores are summarised and illustrated in Fig. 15. This presentation of the paragenetic relations is based upon the scheme proposed by F. Robertson and P. Vanderveer (1952). The salient features of this paragenetic diagram may be summarised as follows:

(a) The numbered points along the circumference of the circle represent the primary hypogene minerals recognised in the ore suite. The numbers are arranged in numerical order with the oldest mineral represented as one.

(b) Minerals that were observed in contact with one another are connected by solid lines.

(c) The head of the arrows point towards the mineral which has been replaced. A dashed arrow head points to replacement relations which are not clearly illustrated by the ore study, and are regarded as tentative.

(d) The role of exsolution is indicated by a separate mineral location outside the circumference of the circle.

The paragenesis of the Klein Letaba ores shows a sequence of mineralisation under decreasing temperature conditions, which links the process of ore deposition to the varied phases of contact metamorphic action discussed in a previous section.

Following closely upon the pneumatolytic phase, arsenical solutions are considered to have induced early crystallisation of the di-arsenide lollingite. These arsenical solutions appear to have become increasingly enriched in sulphur during the alkali-silica metasomatic phase. Concentration and deposition of arsenopyrite, pyrite and pyrrhotite, must have occurred after the alkali-silica metasomatism had effected large scale alteration of the sheared amphibolite host rock, as these hypogene ore minerals extensively replace the silicate gangue.

The close paragenetic relationship between sphalerite, chalcopyrite and gold has already been referred to and it suggests deposition of these minerals under decreasing temperature conditions. F. G. Smith (1943) considers that gold is more soluble than most of the compound metals in alkaline sulphide solutions. The gold would therefore concentrate in the last phases of this alkaline sulphide medium, and deposition would most likely be possible only under relatively low temperature conditions. The deposition of gold, therefore, is considered to be closely related to the low temperature hydrothermal action noted during the last phases of the contact metamorphism described in a previous section.

The enrichment in soda noted in the petrological investigation of the metasomatised zones tends to substantiate the alkali sulphide theory of gold deposition. In addition, mention may here be made, of the association of the gold bearing ores at Klein Letaba with albite-oligoclase felspar, which, according to D. Galacher (1940) appears to be a characteristic feature of gold ores in many parts of the world.

Genetic Classification. The association of the Klein Letaba ores with abundant tourmaline points to their formation under relatively high temperature contact metamorphic conditions, and permits the ores to be classified with reasonable confidence as belonging to the hypothermal class, as defined by W. Lindgren (1933). The prevalence of the diagnostic high temperature sulphide, pyrrhotite, further supports the view that the major hypogene ore constituents recognised in the mine were deposited under hypothermal conditions.

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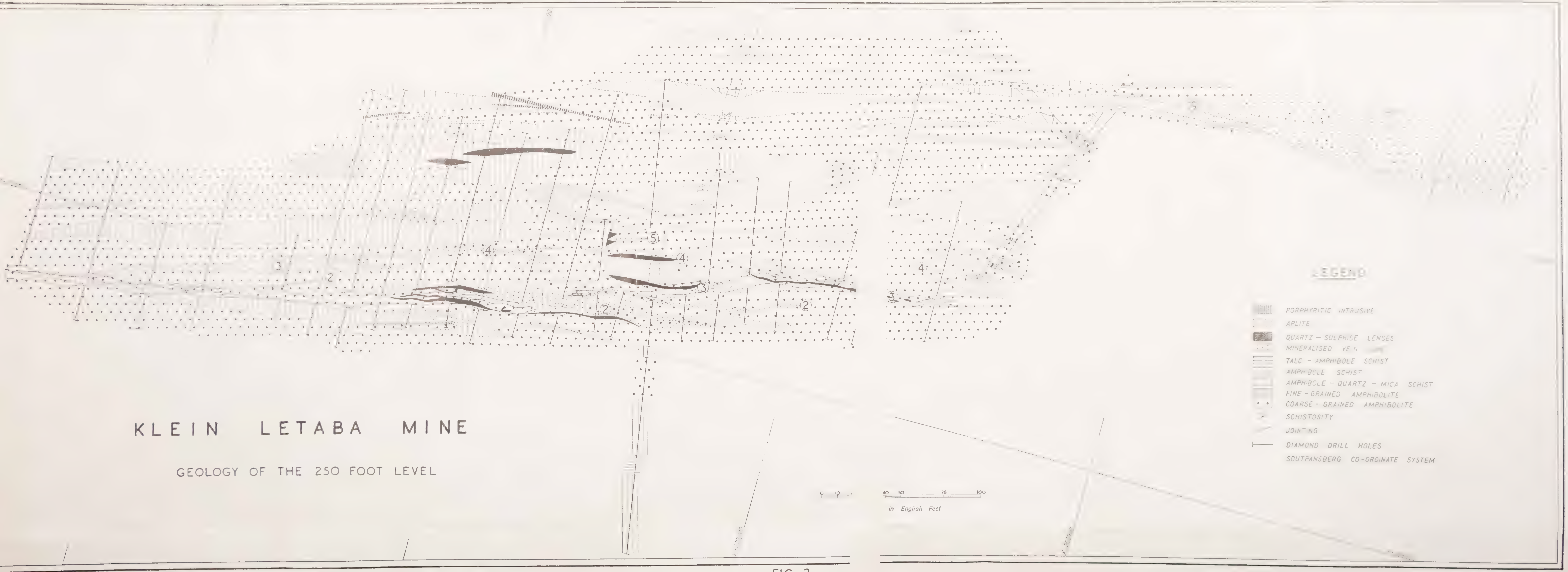
KLEIN LETABA MINE
GEOLOGY OF THE 250 FOOT LEVEL
LEGEND
PORPHYRITIC INTRUSIVE
APLITE
QUARTZ – SULPHIDE LENSES
MINERALISED
TALC – AMPHIBOLE SCHIST
AMPHIBOLE SCHIST
AMPHIBOLE – QUARTZ – MICA SCHIST
FINE – GRAINED AMPHIBOLITE
COARSE – GRAINED AMPHIBOLITE
SCHISTOSITY
DIAMOND DRILL HOLES
SOUTPANSBERG CO-ORDINATE SYSTEM
0 10
40 50 75 100
In English Feet

FIG. 2

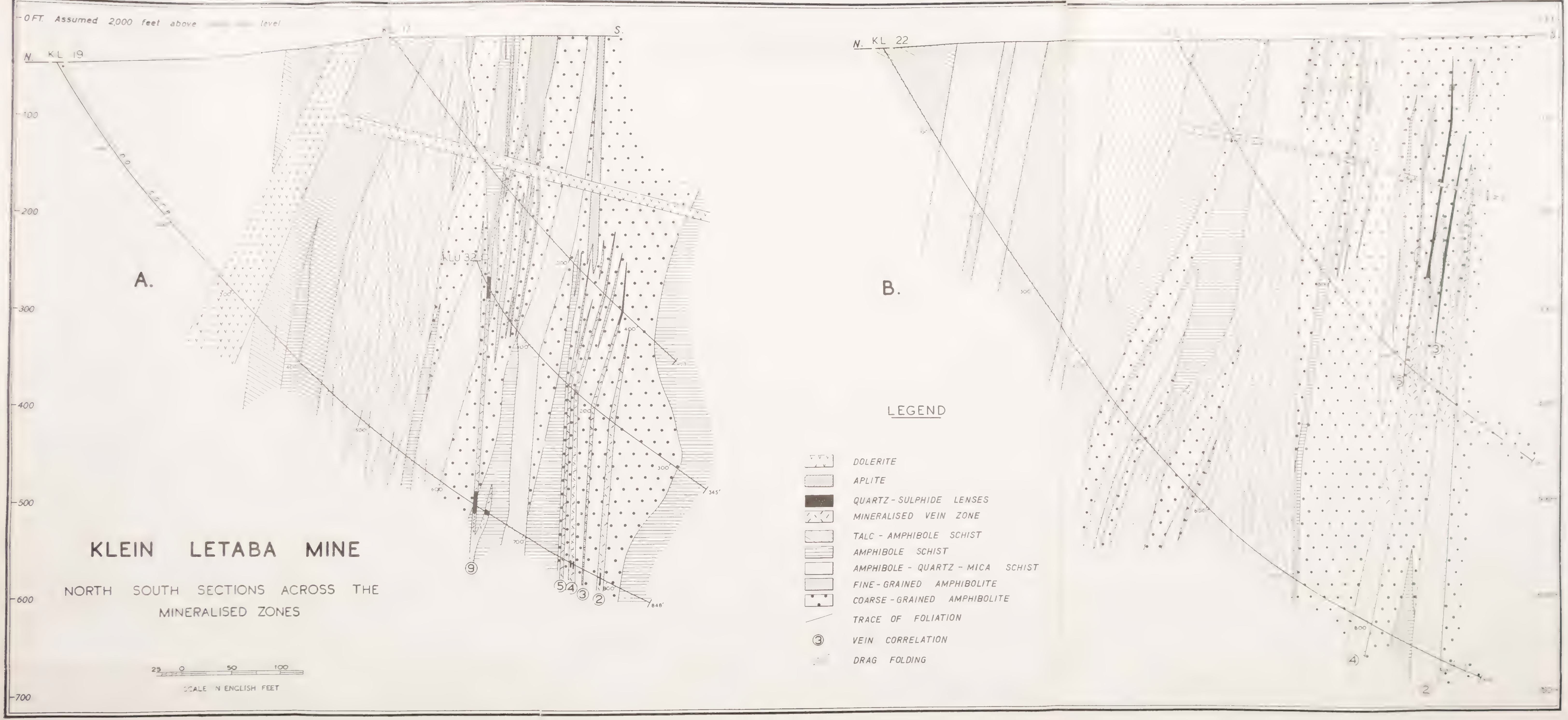
KLEIN LETABA MINE
NORTH SOUTH SECTIONS ACROSS THE MINERALISED ZONES
A.
B.
N. KL 19
S.
N. KL 22
KLU 32
0 FT. Assumed 2,000 feet above
level
100
200
300
400
500
600
700
345'
848'
LEGEND
DOLERITE
APLITE
QUARTZ - SULPHIDE LENSES
MINERALISED VEIN ZONE
TALC - AMPHIBOLE SCHIST
AMPHIBOLE SCHIST
AMPHIBOLE - QUARTZ - MICA SCHIST
FINE - GRAINED AMPHIBOLITE
COARSE - GRAINED AMPHIBOLITE
TRACE OF FOLIATION
VEIN CORRELATION
DRAG FOLDING
25 0 50 100
SCALE IN ENGLISH FEET

FIG. 3

A PETROGRAPHIC STUDY OF THE WATERFALL GORGE PROFILE AT INSIZWA

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(Submitted January, 1957)

ABSTRACT

A detailed petrographic description of a complete succession of rocks over a vertical range of 2,130 feet of the Insizwa intrusion is given. Special emphasis has been laid on rock textures and the nature and orientation of orthopyroxene lamellation as well as clinopyroxene intergrowths. Seventeen new chemical analyses are included and the variation in chemical and mineralogical composition of the pluton from the floor almost to the roof of the intrusion is diagrammatically represented and described. The study suggests that the original magma, the bulk of which was more basic than tholeiite, has been emplaced in at least two successive flows.

CONTENTS

I. INTRODUCTION

The injection of tholeiitic magma into the sediments of the Karroo System during early Jurassic times has given rise to a vast number of intrusive bodies over an area probably exceeding a quarter million square miles. The most prevalent forms displayed by these bodies of "Karroo Dolerite" are discordant dykes and concordant sheets and sills, the latter generally varying in thickness between one and 1,000 feet. Of the sub-horizontal, transgressive bodies, the thickest so far observed is that in the vicinity of Mount Ayliff in East Griqualand, where the huge undulating sheet of the Insizwa-Ingeli-Tonti-Tabankulu intrusion is more than 3,000 feet thick.

Considerable economic interest is attached to this sheet-like body as a result of the occurrence of magmatic nickeliferous ore deposits along certain depressions of its lower contact. Significantly, the igneous rocks of this intrusion are characterised by the most advanced stage of magmatic differentiation of all known occurrences of Karroo Dolerite.

Previous Literature.

The earliest published description of the Insizwa intrusive is by A. L. du Toit (1910). This was followed by publications by W. H. Goodchild (1916) and T. C. Phemister (1924). To date, the most comprehensive work is by D. L. Scholtz (1936), "The Magmatic Nickeliferous Ore Deposits of East Griqualand and Pondoland", in which the previous literature, mentioned above, is summarised. Though primarily concerned with the mineragraphy, distribution and genesis of the ores, Scholtz gives a very complete picture of the general geology and generalised form of the intrusive, as well as a detailed account of the petrography of the lower portions of the intrusive. In summarising the general geology in the next section, the writer has drawn freely from these descriptions and conclusions.

Publications on the petrology of the Karroo dolerites as a whole are numerous, the most recent and comprehensive being that of F. Walker and A. Poldervaart (1949). While no new observations on Insizwa are made in that report, many interesting features of the differentiation and attendant mineral variations of other occurrences of Karroo dolerites are discussed.

General Geology.

The large, isolated mountain masses in the vicinity of Mount Ayliff consist of gabbroic rocks, underlain and, to a very limited extent, capped by sediments of the Lower Beaufort and Upper Ecca Series of the Karroo System. In general these sediments are relatively undisturbed with a gentle dip towards the north-west. Immediately adjacent to the gabbro contacts, the sediments have been intensely metamorphosed to dense hornfels and their dips may show local steepenings along the basal contact amounting in places to 10° or 15°. All the mountain masses, the Insizwa as well as

the Ingeli, Tonti and Tabankulu, are characterised by trough-shaped or basin-like lower margins, which dip inwards at angles ranging from less than 15° to as much as 40°. The study of the pseudo-stratification, cryptic layering and occasional mineral banding (most apparent in the lower parts of some of these masses) indicates a gradual but perceptible flattening of the inwardly directed dip at successively higher horizons within each mass, which may be regarded as being constituted of several superimposed indefinitely bounded, concavo-convex lenses, varying in thickness, lateral distribution and degree of curvature from base to summit.

A. L. du Toit, as well as D. L. Scholtz, believes that the mountain masses constitute part of a single sheet varying in thickness from less than 1,000 feet to more than 3,000 feet, which originally covered an area of not less than 700 square miles, and whose volume exceeded 300 cubic miles. The mountain masses then appear to be the synclinal relics of the original intrusive sheet which possessed markedly undulatory but rudely parallel upper and lower surfaces. The ridges separating these troughs appear to have been somewhat thinner and have been removed by erosion.

Since the elevation of the country from the end of the Mesozoic onwards, erosion removed the overlying strata: most probably a thick pile of Drakensberg lavas, the Stormberg sediments and most of the Beaufort series, the total thickness of which is estimated by D. L. Scholtz as being between 10,000 feet and 16,000 feet. As he is of opinion that no fissures traversed the entire thickness of the overlying sediments and lavas, he deduces *for this closed system a pressure of at least 750 atmospheres, at the time of consolidation.* Underlying the intrusion are to be expected the Ecca Series, Dwyka Series and the Table Mountain Sandstone, with a total thickness estimated to be some 5,000 to 7,000 feet. D. L. Scholtz thinks that the injection took place in at least two, in all probability more, successive and almost synchronous magmatic impulses.

Magmatic differentiation during the cooling of the original intrusive has given rise to three distinct petrological units:

(1) *Roof zone,* or upper acid phase of granitic composition which attains its maximum development in the crests of the arches of the intrusion.

(2) *Central zone,* consisting primarily of gabbro and constituting the major portion of the existing mountain masses.

(3) *Basal zone,* characterised by picrites and olivine-rich hyperites which have been preserved from erosion in the troughs underlying the main mountain masses.

Scholtz believes this division is due chiefly to gravitative differentiation with the settling of olivine into the basal zone. He also maintains that the metallic sulphides separated as an immiscible liquid from the silicate magma at an early stage of its cooling history. Under the influence of gravity, this liquid fraction accumulated in the lower portions of the basal zone, where it solidified after the crystallisation of the enclosing silicates.

Present Investigation.

Early in 1950, a number of old mine workings in Waterfall Gorge, along the lower contact of the differentiated sheet on the southern flank

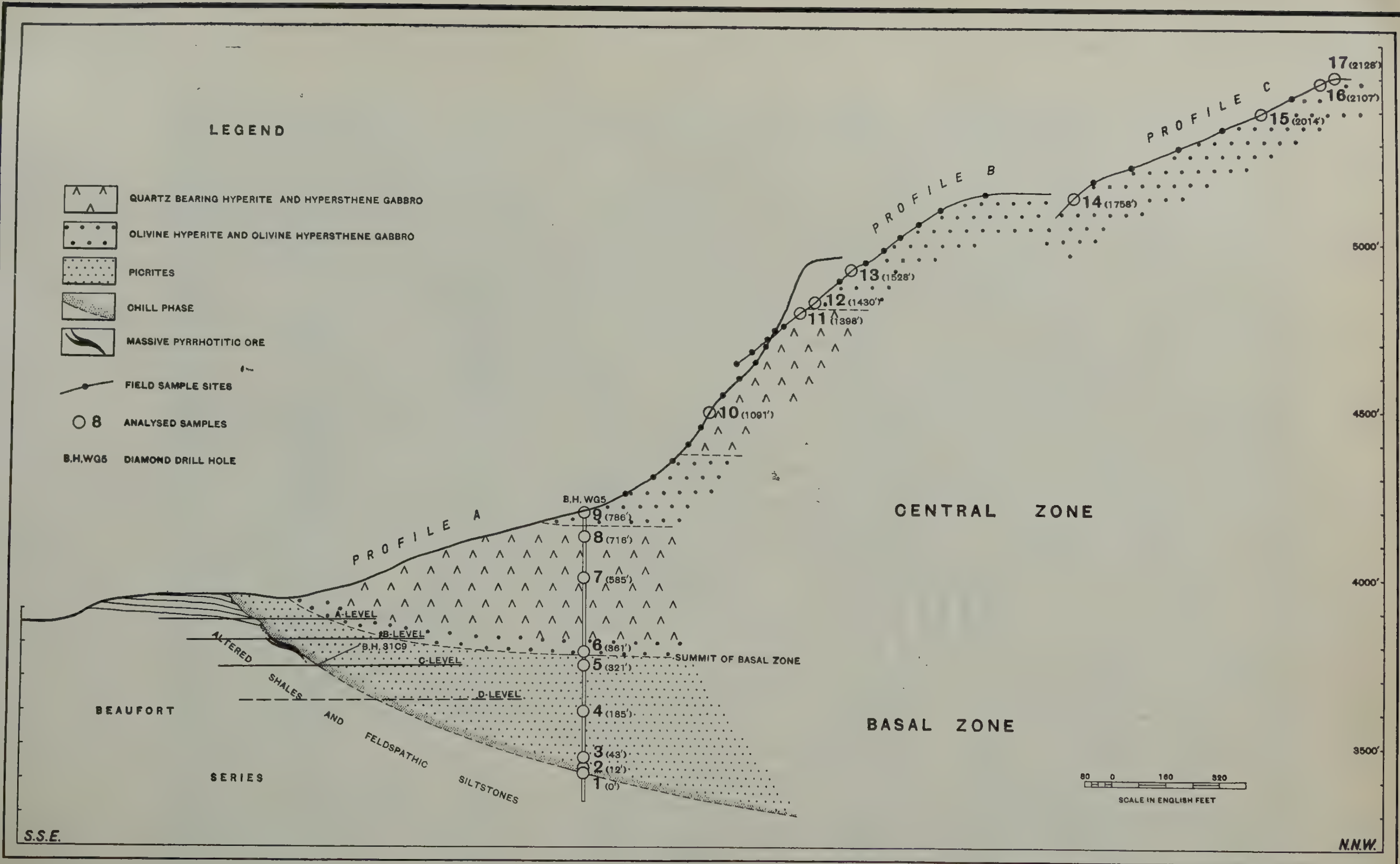

FIG. I

Fig. 1. Waterfall Gorge section of the Insizwa.

of the Insizwa Range, were de-watered, and an extensive prospecting programme was initiated in this vicinity by the Nickel Corporation of Africa, Limited.

In August, 1951 the core of an 800-foot vertical borehole, which had been sunk through the basal zone and part of the central zone for prospecting purposes some ten years before, was made available for study by the director of the Geological Survey of the Union of South Africa. This material offered an ideal opportunity for studying the differentiation and mineralogical variation across the lower portions of the intrusive at accurately measured heights above the floor. As such a precise height-control was not available to D. L. Scholtz during his sampling, the present investigation was undertaken to supplement the previous work with greater detail. It was also decided to extend the investigation to horizons above those represented in the core by taking samples on the surface slope above the level of the collar of the borehole.

Through the courtesy of the Nickel Corporation of Africa, Ltd., a dip section and topographic profile (Figure 1) through the borehole, W.G. 5, showing the probable degree of flattening of the basal contact as revealed by sub-surface prospecting operations at Waterfall Gorge was placed at the disposal of the writer. Eleven additional samples, covering a vertical range of 550 feet, were taken along this profile, but the abrupt steepening of the topography necessitated offsetting the remaining sample profiles up to 500 feet westwards. The entire sampled profile therefore represents a vertical range of 2,130 feet, and extends from the floor of the intrusive to the highest point of the Insizwa Range in this vicinity, a very short distance below the upper contact of the sheet. The offset profiles are indicated in Figure 1, by means of the profile lines B and C. After a preliminary investigation of all the samples, 17 were selected for detailed investigation and chemical analysis. Their modes and norms are listed in Table I.

II. PETROGRAPHY

In the profile under consideration, a complete suite of rock specimens of the Basal Zone and the Central Zone were available for study.

All the rocks in this profile, except the lower chilled selvage, are medium-grained, and range in composition from hyperites and hypersthene gabbros free of olivine to picrites with more than 60% olivine. The rock classification used is that which was adopted by D. L. Scholtz (1936), which he based on the classifications of A. Holmes (1928) and S. J. Shand (1927).

Olivine Gabbro Series $>$ 30% Felspar, $<$ 35% Olivine	Picrite Series 5-30% Feldspar, $>$ 35% Olivine
Olivine Norite (orthopyroxene)	Picro-norite (orthopyroxene)
Olivine Hyperite (ortho$>$ clino)	Hypro-picrite (ortho $>$ clino)
Olivine Hypersthene Gabbro (ortho $<$clino)	Hypersthene-picrite (ortho $<$ clino)
Olivine Gabbro (clinopyroxene)	Picrite (clinopyroxene)

In this study the sub-divisions of the picrites have been omitted.

The dominant feature in the succession of the rocks is the repetition of the olivine-bearing zones, separated by olivine-free zones (see Diagram I). The Basal Zone, 355 feet thick at W.G. 5, in the section under consideration, forms the main portion of the lowest olivine-bearing zone, which attains a thickness of 410 feet. It is overlain by a 320-foot thick olivine-free zone. The second, the "central", olivine-bearing zone of 225 feet is overlain by the central quartz-bearing zone of 450-foot thickness. The third, the upper olivine zone, attains a thickness of at least 713 feet, and extends to the summit of the range in this section.

The olivine content at the top of the section decreases suddenly to less than 5% in the last sample. This may indicate that here the top of this thick upper olivine-bearing zone is reached. A thin section of a sample from the summit of the range, some hundred feet to the west of the profile under investigation, proved indeed to be typical of the olivine-free zones.

If an olivine-free zone had been present above the uppermost concentration of olivine in the profile under consideration, it must have been thin, because the overlying hornstones are found at a lower topographic altitude further to the north.

BASAL ZONE.

The Basal Zone consists of very dark greenish grey, medium-grained rocks.

Its upper limit is located just below the 361-foot level (sample no. 6), where, within ten feet, the olivine content is found to drop from 45% to less than 25%. Here are also sudden changes in mineral composition and a significant change in the habit of the plagioclase crystals from large poikilitic plates to normal subhedral laths. The demarcation of this limit corresponds to that established by Scholtz, although the troctolite, which he uses as a convenient upper marker for the Basal Zone elsewhere, is not developed in this section. In both cases the variation in the olivine content, however, distinguishes the rocks of the Basal Zone containing more than 25% olivine, from those of the Central Zone which hold less than 25% olivine.

In this section the Basal Zone is found to contain three petrographic sub-divisions grading imperceptibly into one another:

(iii) Picrites
(ii) Olivine-rich hyperites
(i) Chill phase.

(i) *Chill Phase,* occurring along the contact of the Basal Zone and the underlying hornfels, is composed of a dense, nearly black, very fine-grained rock.

In this section it is found to be composed largely of dendritic globular aggregates of pyroxene about 1.5 mm. long, but sometimes even twice as long, in a groundmass of very fine-grained plagioclase forming a characteristic radiating texture (Plate I, pm. 1).

One foot above the contact the plagioclase laths attain lengths of 0.5 to 1.3 mm. The cores of the crystals are clouded with dust-like inclusions; the rims are usually clear. Some alkali feldspar and quartz are associated with these more acid rims, and fill the spaces between the laths.

Highly altered olivines 0.7 mm. in diameter in the form of aggregates

of brownish bowlingite and green chlorite are very rare. Biotite on the other hand is comparatively abundant in irregular flakes. Some of these flakes are associated with the opaque minerals, forming rims around small sulphide pellets and irregular ilmenite rods. The presence of octahedra of magnetite and patches of sulphide (though rare) is significant, since they undoubtedly represent primary constituents of the original intrusive magma. Fine apatite needles, often parallel to one another, are found in the clear rims of the plagioclase crystals, as well as in interstitial alkali feldspar and quartz and biotite.

Thin serpentine veins are present throughout the intrusion, cutting both the ore pellets and the biotite. In the chill phase, close to the hornfels contact, these veins also contain calcite, while tiny calcite grains are found sporadically distributed throughout the rocks. The serpentine and calcite are regarded as deuteric alteration products of olivine and basic plagioclase respectively.

(ii) *Basal olivine hyperite* forms a layer, roughly 10 feet thick merging gradually with the chill phase below and the picrite above. The rock is composed essentially of plagioclase laths, pyroxene and olivine crystals in roughly equal proportions, orthorombic pyroxene being more abundant than the monoclinic variety.

Three feet above the contact the olivine is usually fresh. Both the size and abundance of the olivine are found to increase at progressively higher levels until the overlying picrite, with an olivine content exceeding 35% by volume and an average grain size of about 0.7 mm., is reached. The olivines often possess ragged and embayed margins.

The pyroxenes show a textural variation across the olivine hyperite. In the lower portion, stringers of pyroxene contain differently orientated grains of augite intergrown with and surrounded by orthopyroxene simulating a coarser modification of the pyroxene habit which characterises the chill phase. At slightly higher levels, however, the pyroxene stringers become rarer and hypersthene and augite tend to occur in isolated grains instead of aggregates. Some pigeonite is associated with the augite forming the cores of augite crystals or constituting part of augite clusters (Plate I, pm. 3).

In general the plagioclase laths too become larger at higher levels and their clouded cores become less conspicuous, while their clear rims become wider.

Among the accessory minerals, magnetite, associated with the olivine and pyroxenes, is more abundant in the olivine hyperite than in the chill phase. Minute sulphide pellets occur together with some ilmenite and apatite needles, and are embedded in the biotite and in the interstitial acid plagioclase and alkali feldspar.

(iii) *Picrite* comprises the dominant rock type in the Basal Zone. In this section it extends from 15 to 355 feet above the basal contact.

The olivine content of these rocks exceeds 35% by volume and may be as great as 62% between the 185-foot and 321-foot levels. Despite their abundance, the growth of the olivine crystals is seen to have been relatively free from mutual interference. Sometimes a number of crystals are grouped together in clusters, in which the individuals join each other along prism faces. Contacts between olivine and plagioclase crystals usually show the

preserved crystal faces of the olivine, but the olivine crystals are always well-rounded and resorbed when in contact with pyroxene.

The plagioclase crystals in the picrite possess a poikilitic habit, large plagioclase oikocrysts up to 3 mm. containing abundant olivine chadacrysts. These plagioclases, especially towards the top of this zone, are often in sub-parallel orientation forming patches of more than 1 cm. in length. This feldspar is strongly zoned. The pyroxenes, both bronzitic and augitic, also poikilitically envelop olivine but the chadacrysts are fewer in number and rounded or embayed.

Of the opaque minerals, magnetite octahedrons are the most abundant. They are generally enclosed in olivine or pyroxene but when in plagioclase they remain in contact with the olivine crystals. Sulphides occur in irregular grains squeezed between the other minerals and surrounded by irregular collars of biotite flakes. Apatite is present as extremely fine needles embedded in the late plagioclase and alkali feldspar.

CENTRAL ZONE.

The rocks of the Central Zone are medium-grained and somewhat lighter in colour than those of the Basal Zone. The following divisions are recognised on the basis of the variation in the proportions of the principal minerals present:

(vi) Upper olivine hyperite
(v) Upper olivine hypersthene gabbro
(iv) Central quartz-bearing hypersthene gabbro
(iii) Central olivine hyperite and olivine hypersthene gabbro
(ii) Lower quartz-bearing hyperite and hypersthene gabbro
(i) Lower olivine hyperite.

The olivine in these horizons seldom exceeds 15%. In contrast the plagioclase content usually varies between 50% and 60%, while the pyroxenes comprise between 25% and 40% of the rock. The only exceptions are the olivine-rich horizons at the 786-foot level and the plagioclase-rich horizon at the 841-foot level, both in the central sub-zones. The mafic minerals tend to form clusters and this feature results in the slightly mottled appearance of the rocks.

The sharp decrease in olivine content in comparison with that in the Basal Zone, is accompanied by a change in the textural relation between the plagioclase and the mafic minerals. In the Central Zone the plagioclase forms small laths which interfered with the growth of the olivine crystals and are enclosed in both olivine and pyroxene crystals. Therefore it seems that these mafic minerals of the Central Zone crystallised only after basic plagioclase laths were present around them.

The different rock types may now be described individually in ascending order.

(i) *Lower olivine hyperite*

In this sub-zone it appears that the transition from the Basal Zone to the lower quartz hyperite is discontinuous. The change in textural relations between the plagioclase and the olivine and pyroxenes between the 351-foot

and 361-foot levels has already been described. But between the 366-foot and 376-foot levels another discontinuity has been observed. Here the olivine content, after a sharp decrease in the lower 10 feet of the Central Zone, is absent at the 371-foot level. But it is present again a foot higher in the profile, attaining a maximum amount of about 30% at the 376-foot level before it gradually diminishes in volume percentage and disappears near the 411-foot level. This distribution is again accompanied by distinct textural variations in the plagioclase as well as in the pyroxenes.

Both the pyroxenes retain their poikilitic habit at the 361-foot and 366-foot levels. Here the augite, as well as the orthopyroxene, forms large single crystals enclosing resorbed olivines and connected networks of plagioclase laths.

But in the olivine-free horizon at the 371-foot level, the orthopyroxene forms small prismatic crystals, often rounded and in clusters, while irregular grains are enclosed in the augite. The augite is also more finely grained than in the underlying horizons, but it still crystallises in anhedral ophitic grains.

In the overlying olivine-bearing horizons the small plagioclase laths tend to form clusters, and are accompanied by a generation of larger subhedral plagioclase laths enclosing some of the smaller laths in their mantles. The orthopyroxenes are somewhat larger, and enclose plagioclase laths sub-ophitically, but their prismatic cores are free from plagioclase. Where the olivine reaches its maximum concentration, the prismatic habit of the orthopyroxene is less pronounced and equidimensional anhedral grains appear to have been formed instead. The augite retains its ophitic habit throughout this sub-zone.

The accessory minerals, notably sulphides, biotite and alkali feldspar, are relatively abundant, but magnetite is rarely present. Apatite is scarce and forms relatively stout, short prisms.

(ii) *Lower quartz-bearing hyperite and hypersthene gabbro* is approximately 320 feet thick in this section. The disappearance of the olivine is accompanied by the appearance of minor amounts of hornblende associated with the mafic minerals and quartz, which is mostly micrographically intergrown with alkali feldspar. Apatite, biotite and sulphides occur in very limited quantities.

The upper half of the lower olivine hyperite grades imperceptibly into the quartz hyperite, without textural variations in the plagioclase and pyroxenes. The orthopyroxenes are strongly zonal, becoming more iron-rich towards their margins. The amount of hypersthene exceeds that of augite in the lower portion of this sub-zone, but the relationship is reversed in the upper levels. This is accompanied by a general textural variation across this sub-zone: in the lower portion, anhedral sub-ophitic hypersthene crystals possess clear prismatic cores, while their margins contain fine lamellae, blebs and rods of augitic material. From the 495-foot level upwards, however, the hypersthene crystals become smaller, are free of augitic "intergrowths" and possess prismatic habit. The small, rounded hypersthene crystals as also seen in and above the 371-foot level, become more abundant upwards giving these rocks a comparatively more finely grained appearance. These small orthopyroxene grains usually occur evenly distributed between plagioclase laths, but some are enclosed in sub-ophitic augite. The anhedral sub-ophitic

augite crystals also tend to become smaller upwards and often form clusters of interlocking grains.

Within the last 20 feet of this olivine-free sub-zone, a distinct directive structure is developed: elongated laths of plagioclase and anhedral prismatic crystals of hypersthene and augite are arranged along sub-horizontal planes. Consequently the (100) twinning plane of augite, which is parallel to the greatest dimensions of this mineral, is more frequently observed in vertical than in horizontal thin sections. However, no lineation within the horizontal plane could be detected (see Pl. I, pm. 2).

(iii) *Central olivine hyperite and central olivine hypersthene gabbro*

These olivine-bearing rocks form a layer roughly 225 feet thick towards the middle of the Central Zone. They are the lowest rocks in this section, in which the augite has preceded the orthopyroxene in the crystallisation sequence to such an extent that clusters of interlocking grains of augite are enveloped and penetrated by orthopyroxene.

Olivine is often resorbed when bordered by either of the pyroxenes: when in contact with plagioclase, kelyphitic structures are formed. The maximum amount of olivine is 33% and has been found near the base of this sub-zone at the 786-foot level, while elsewhere not more than about 10% olivine is present. Such marked increases in the olivine content are seen to be accompanied by some variation in texture and relative proportions of orthopyroxene and augite. In this sub-zone both pyroxenes appear to have crystallised as separate anhedral crystals, rather than as poikilitic aggregates, while the ratio of orthopyroxene to augite shows a marked increase. Though the olivine in this horizon is typically euhedral, grains enclosed in plagioclase and pyroxene appear to have been strongly resorbed. No kelyphitic borders have been formed.

Of the remaining 50 feet of this sub-zone below the olivine-rich horizon, one sample is available from the 751-foot level. Here the olivine occurs as small rounded grains in clusters; kelyphitic borders are again well developed. Here too, and in some horizons immediately overlying the olivine hyperite, the orthopyroxene is only occasionally pseudo-poikilitic, usually occurring as isolated large ophitic grains not directly associated with the olivine and augite clusters.

The accessory minerals of the olivine hyperite include sulphides, magnetite and probably some ilmenite, biotite, apatite and some interstitial alkali feldspar. Hornblende and quartz are again absent.

(iv) *Central quartz-bearing hypersthene gabbro*

In this second olivine-free sub-zone, the amount of augite reaches a maximum for the section under consideration. Together with plagioclase it constitutes from 85% to 90% of the volume of the rocks. The augite throughout this sub-zone forms clusters of interlocking anhedral grains, which may envelop some plagioclase.

The plagioclase laths generally show the complex zoning features typical of the plagioclase of most horizons. But in the upper 130 feet of this sub-zone they do not possess the typical oscillatory zoned cores. This feature is accompanied by an increase in grain size: the maximum length attained by the crystals is usually 3.5 mm. instead of 2 mm.

The hypersthene appears to possess two distinct habits in this sub-zone. On the one hand, the orthopyroxene occurs as large ophitic to pseudo-poikilitic crystals enclosing rounded and embayed augite grains of varying orientation, similar to the orthopyroxene in the underlying rock-type. In addition these large crystals may possess patches in which orthopyroxene and augite are graphically intergrown (see Fig. 4 (b, e, f)). On the other hand, smaller hypersthene crystals may occur as individual grains, each closely associated with one crystal of augite. Augite intergrowths are also present in these hypersthenes, but they are more regular than in the poikilitic orthopyroxenes. Hypersthene possessing this habit is confined to the upper 130 feet of this sub-zone (Fig. 4 (d)).

Quartz and alkali feldspar, intergrown micrographically, become quite conspicuous in this sub-zone, and some hornblende is present. The other normal accessory minerals, biotite, ore and apatite, are extremely rare.

(v) *Upper olivine hypersthene gabbro*

The average grain-size of the lower 100 feet of this olivine hypersthene gabbro is somewhat less than that of the remaining rocks of the Central Zone. The remaining 575 feet possess a coarser texture, however, comparable with the rest of the intrusive.

The olivine crystals tend to form clusters or strings. They are the most Fe-rich varieties of this section (Fa_{31}-Fa_{34}) and appear to have grown against plagioclase laths, sometimes even enclosing some small crystals of the feldspar.

Clusters of interlocking augite grains are often found to enclose plagioclase laths, while pseudo-poikilitic orthopyroxene includes rounded olivine as well as rounded augite grains and often forms very thin rims around these chadacrysts. Graphic intergrowths between these two minerals are sometimes present in the lower half of this sub-zone.

The accessory minerals apatite, ore and interstitial alkali feldspar, are present in only minute quantities.

Towards the centre of this zone, represented by sample number 14, there is a local increase in the amount of olivine accompanied by a slight change in the habit of the pyroxene, as has been observed in sample number 9 of the central olivine-bearing zone. Here the augite appears in larger crystals enclosing some olivine, while orthopyroxene forms some relatively large patches which ophitically envelop plagioclase but are free of either olivine or augite. The two last named minerals may, however, be poikilitically surrounded by optically continuous coronas of hypersthene.

(vi) *Upper olivine hyperite*

The highest specimen collected along this section, at a vertical height of 2,128 feet above the lower contact of the intrusive, is a fine-grained rock with a coarsely mottled appearance. It is an olivine hyperite differing from the underlying olivine hypersthene gabbro by containing notably less olivine and more orthopyroxene.

The dark clots in this rock owe their origin to concentrations of olivine and augite clusters, while the lighter portions contain more plagioclase laths and large ophitic orthopyroxenes.

III. MINERALOGY

Olivine.

Olivine is clearly the first mafic mineral to have started crystallising in this intrusion. In the Basal Zone one olivine crystal was found enclosing a grain of bronzite. But as the olivine in general is strongly resorbed the "inclusion" may be the result of the fortuitous sectioning of a strongly resorbed olivine.

As previously mentioned, the olivine is distributed in three distinct zones: (1) in the Basal Zone and 50 foot above it; (2) in the central olivine-bearing zone between the 745-foot and 950-foot levels; (3) in the upper olivine-bearing zone from the 1,415-foot level upwards. (See Diagram I).

In the picrite the maximum amount of about 63% olivine is reached at a height of 185 feet above the contact and tends to remain constant for the following 150 feet, after which it diminishes rapidly. In the central olivine-bearing zone the maximum amount of 33% is found only 45 feet above the lower margin of this sub-zone, but in the overlying 150 feet of this horizon it does not constitute more than 10% by volume of the rock. In the upper olivine hypersthene gabbro the maximum concentration of olivine is found, at a height of 1,758 feet, in the middle of this sub-zone. This maximum is the lowest of the three and amounts to 26.5%. As will be seen, the distribution of the olivine is closely connected with the composition not only of the olivine itself, but also of the augites and orthopyroxenes (see Diagram I). In addition the distribution of the olivine also appears to be related to textural variations exhibited by these mafic minerals and by the plagioclase. Euhedral olivines seem to be restricted to those crystals which are bordered by plagioclase in the zones with more than 25%-30% olivine, i.e. the picrite and central olivine hyperite. When they are in contact with pyroxenes, however, they always show anhedral or resorbed margins. In the horizons with less than 25%-30% olivines, the olivines are feebly moulded on the plagioclase laths and may partly or completely enclose some small laths of that mineral.

The olivine grains are usually equidimensional or only slightly elongated along the *c*-crystallographic axis, and under the circumstances no preferred orientations can be expected. The average length of the crystals is about 1.5 mm., but the grain size varies greatly in every thin section. The largest grains were found in the olivine hyperites *especially above the picrite.* Here crystals may attain lengths of 2.5 mm. or more. One strongly resorbed crystal 5 mm. in diameter was found in the rock 20 feet above the picrite. In the picrite itself the maximum size seldom exceeds 2 mm., and on descending in the succession, the average size decreases very slightly down to 43 feet above the contact. Below this level the grain size diminishes rapidly, the maximum size in the chill phase being only 0.7 mm. In the central olivine-bearing zone the largest crystal measured (in sample number 9) also has a length of 2.5 mm. In the upper olivine-bearing sub-zone there is a gradual but marked increase in the maximum size from 1.3 mm. in sample number 12 to 2.3 mm. in sample number 16.

The relation between the amount of olivine present and the variation in the size of the optic axial angle (implying the composition of the olivines)

is demonstrated in Diagram I. The composition of the olivine was determined by optic axial angle measurements on the universal stage. The angles range from $2V_\gamma = 89^\circ$ in the picrite to $2V_\alpha = 81^\circ$ in the upper olivine-bearing zone and correspond to 11% Fa and 34% Fa (Winchell, 1933) respectively.

It is noteworthy that interpenetration twins of olivine occur in the Insizwa rocks. An example is illustrated in Plate I, pm. 5. Examination on the universal stage revealed that the twinning axis is perpendicular to (011). These twinned olivines are extremely rare and were only found in the picrite 85 feet above the basal contact and in the central olivine hyperite.

In some rocks the olivines show kelyphitic borders when in contact with plagioclase. Each border probably consists of a narrow rim of anthophyllite followed by garnet with a myrmekitic structure. These secondary reaction rims have been found only in rocks of the central and lower olivine-bearing sub-zones (containing less than 20% olivine). They are best developed in the central olivine-bearing sub-zone. In the upper olivine-bearing zone, however, the kelyphitic borders are absent or insignificant.

A normal zoning with increasing Fa-content towards the crystal margins was noticed in some olivines in the middle of the picrite and near the bottom of the upper olivine-bearing zones.

The minute dendrites of magnetite in the olivines, as described by A. L. du Toit (1910), were found to be present in the Basal olivine hyperite below the picrite.

Plagioclase.

The plagioclase throughout the section is rather uniform in appearance. It usually crystallises as subhedral, strongly zoned laths which are seldom longer than 2 mm. In the picrite, however, the plagioclase is present as large poikilitic zonal crystals enclosing the olivine.

Small plagioclase laths are often enclosed in the less Mg-rich olivines and pyroxenes, while the more acid mantles of the adjacent crystals are moulded on the same mafic minerals. This indicates that in most horizons the period of crystallisation of plagioclase was longer than that of the other minerals. In the Basal Zone, however, the plagioclase is always moulded on the olivine, while the pyroxenes are moulded on and wedged between the plagioclase laths, indicating that this plagioclase crystallised after the olivine and mainly prior to the pyroxenes. At the top of the picrite the poikilitic plagioclase envelopes euhedral olivine chadacrysts in contrast to the poikilitic pyroxenes, which enclose more strongly resorbed olivine chadacrysts, fewer in number and with highly embayed outlines.

Walker (1940) believes that the large size of plagioclase is in some way connected with concentrations of orthosilicates. The poikilitic habit of the plagioclase in the picrites of Karroo dolerites is ascribed by Walker and Poldervaart (1949) to crystallisation in a quiescent magma. They concluded therefore that these picrites have been formed in situ by crystal accumulation and not by differentiation prior to the emplacement.

While agreeing with the general conclusions of these authors the writer believes that the large size and poikilitic habit of the essential mineral constituents other than olivine should be attributed primarily to the quiescent state of the interstitial magma, in which volatiles derived from the lower

levels have become concentrated owing to the seating action of overlying heavy accumulations of olivine and the immediately overlying early and more basic plagioclase. This conclusion is substantiated by the presence of biotite, alkali feldspar, etc., and the lower total anorthite content of the plagioclase in the picrite as well as in other olivine-rich rocks.

The modes (Diagram I) illustrate the antipathetic relationship between the volumetric proportions of plagioclase and olivine in the picrite, the central olivine hyperite and the olivine-rich horizon of the upper olivine hypersthene gabbro. Near the top of the intrusion a similar though far less prominent quantitative relationship also appears to exist between augite and plagioclase.

The composition of the plagioclase was determined by means of the universal stage. As the crystals throughout this section show strong complex zoning, the average compositions given are approximate. In Diagram I the average composition as well as the range in composition revealed by zoning are plotted against the height. A remarkable feature of these curves is that they show no general regular variation in the An-content of the plagioclase across the section, and that there appears to be no obvious relation between the composition of the plagioclase and the variation in the amount of olivine.

Throughout this section the plagioclase crystals are intricately zoned, commonly possessing oscillatory zoned and resorbed cores and normal but discontinuously zoned mantles. The oscillatory zoning is seen, however, only in a thin section whose plane fortuitously intersects the zoned core of the particular feldspar under consideration. In other words, oscillatory zoning will only be apparent in suitably orientated sections passing through the crystal cores. Examination reveals that oscillatory zoning is present in plagioclase crystals of all horizons, with the exception of the relative large plagioclase laths between the 1,270-foot and 1,400-foot levels (upper half of central quartz hypersthene gabbro). It is also found that the oscillatory zoned cores commonly contain some plagioclase of a variety more calcic than 80% An. The most basic plagioclase does not necessarily characterise the innermost core but more often forms one of the surrounding zones. The innermost core may be irregular and suggestive of resorption, or it may be idiomorphic; it usually has the same composition as the mantles which repair the corroded oscillatory zoned cores (see Plate I, pm. 6). This is not an uncommon feature in Karroo dolerites, and was also observed by A. M. J. de Swart and L. J. Murray (1944) in the plagioclase of the Paardekop sill.

The mantles of the feldspars are always normally zoned. The zoning though often continuous may show discontinuities, sometimes with resorption, between the composition ranges ± 75-65% An and ± 60-65% or 40% An.

In the Basal Zone the plagioclase cores do not contain more than 80% An, and the most acid margins or patches often hold less than 50% anorthite. In the large poikilitic crystals at the top of the picrite small patches approximately An_{90} Ab_{10} in composition may sometimes be seen.

Immediately above the Basal Zone, in the lower 200 feet of the Central Zone, the plagioclase laths show a still wider range in composition, accompanied by well developed complex zoning. The small laths enclosed in the pyroxenes and the cores of very large crystals are the most basic and attain a composition approximately 90% An. Those in the pyroxenes are zoned with rims of 70-75% An, but those crystals which were not protected

against later resorption, and the small laths partly or completely enclosed in the mantles of the large plagioclase laths, have cores of about 70% An and rims of 40-50% An.

In the overlying sub-zones cores of 90% An are very rare. In the lower quartz hypersthene gabbro, the variation in composition is small, the maximum range being 83-60% An from core to mantle. Hence it is the most basic plagioclase found in this section of the intrusion. The somewhat larger crystals of the central olivine hyperite 70 feet higher are slightly less calcic.

The composition of the feldspars remains fairly constant in the overlying sub-zones, but in the upper 200 feet of the quartz hypersthene gabbro, the oscillatory zoned cores of 85% An are again absent, while the rims contain about 50% An. These crystals show simple reversed zones and resorbed cores of about 70-75% An and are probably the most sodic plagioclases of the profile. Those of the upper olivine-bearing sub-zones are, however, nearly as basic as the plagioclase of the lower quartz hypersthene gabbro. Their compositions range from 85% An for the core to 60% for the rim. In short, it therefore seems that the most basic plagioclase occurs in the lower quartz-bearing zone and in the upper olivine-bearing zone, while the most sodic plagioclase is present in the Basal Zone and just below the upper olivine-bearing zone.

Both normal and complex twins are present in the plagioclase crystals. In the large poikilitic crystals, Albite and Pericline twins are approximately equally abundant. Albite twins are, however, the most numerous in other crystals, followed by twins according to the Carlsbad law. Albite-Ala, Acline, Baveno, Ala and Manebach twins are also present but are rare, being associated with the most calcic plagioclases and occurring most frequently in the small laths enclosed in the ophitic pyroxenes of the lower olivine and quartz hyperites.

Pyroxenes.

Both clinopyroxenes and orthopyroxenes are present at all horizons throughout the section of the intrusion under consideration.

The clinopyroxene is usually an augite with a very narrow composition range, varying from about $Wo_{42}En_{48}Fs_{10}$ in the picrite to about $Wo_{37}En_{44}Fs_{19}$ in the central quartz hypersthene gabbro. Pigeonite is present only in the lowest 30 feet of the intrusion. The orthopyroxene has a wide range in composition varying from bronzite with 17% Fs in the picrite to hypersthene with 50% Fs in the central quartz hypersthene gabbro.

(i) *Pigeonite.*

The pigeonite, found near the basal contact, forms irregular grains, often in clusters similar to augite with which it is associated. Pigeonite is also found as cores within augite crystals (Plate I, pm. 3). Under crossed nicols the mineral is characterised by a vermicular texture on a very small scale, sub-parallel to (100). No relation between the optic orientation of the pigeonite and the enveloping augite could be found.

Determinations by means of the Universal Stage yielded the following optical constants:

optic plane // (010), $\gamma - \alpha = 0.018$ (core) — 0.020 (mantle)
$2V\gamma = 23°$ (110) $\wedge$ (1$\bar{1}$0) = 88°
$\gamma/c = 41° \pm 2°$

These constants indicate a composition of $Wo_{15}En_{60}Fs_{25}$ (Winchell, 1951).

A. Poldervaart and H. H. Hess (1951) define a pigeonite as "a lime-poor clinopyroxene (Wo_{5-15} percent) with optic axial angles invariable below 30° (usually below 25°), whether in the plane (010) or, as is commonly the case, perpendicular to that plane. H. Kuno (1955) describes three "sub-calcic augites" from basalts. These are strongly, but continuously zoned clinopyroxenes (the optic axial angle ranges from about 10° for the core to about 40° for the margin) in which the values between $2V\gamma = 20°$ and $2V\gamma = 35°$ are quite common, and in which the optic plane is always parallel to the clinopinacoid. The lime-poor clinopyroxene near the base of the Insizwa intrusion may therefore also be regarded as a sub-calcic augite.

The presence of this mineral near the chill phase corroborates the suggestion of D. L. Scholtz (1936) that the pigeonite originated as a result of rapid cooling. A. B. Edwards (1942) suggests a similar origin when he describes a "metastable form of pigeonite" near the basal contact of the Tasmanian dolerites formed "under conditions of rapid crystallisation". It is interesting to note that in the last named intrusions, the composition of the pigeonite corresponds to the mean composition of all the pyroxenes in the dolerites. Similarly, the mean normative composition of the pyroxenes in the chill phase at Insizwa agrees exactly with that of the pigeonite as deduced from its optical properties.

The existence of one pyroxene with a composition intermediate between the orthorombic pyroxenes and augites in "the two pyroxene field" (H. H. Hess, 1941) is also ascribed by H. Kuno (1955) to rapid cooling. Kuno shows that at a high temperature small proportions of Fe^{3+} enter the pyroxene structure in the tetrahedral position. This would prevent the strain caused by the Ca^{2+} ions in the structure. When during rapid cooling the Fe^{3+} ions remain in the tetrahedral position, the structure would be able to accommodate Ca^{2+}, Mg^{2+} and Fe^{2+} in an extended range of proportions. In this case only one pyroxene with an intermediate Ca-content (a sub-calcic augite) would exist instead of two varieties, one a calcium-rich and the other a lime-poor pyroxene.

(ii) *Augite.*

The distribution of the augite shows a marked increase in volume percentage from the margins of the intrusion towards the middle of the central quartz-bearing zones. Marked fluctuations in the distribution of the augite correspond to antipathetic variations in the amount of olivine present. The lesser fluctuations correspond to antipathetic variations in plagioclase content.

The composition and the texture of the augite is influenced mainly by the amount of olivine present. The texture, however, depends also on the position in the section.

Usually the augite crystallises as anhedral grains, about 1-2 mm. in diameter. It then tends to form clusters of interlocking grains moulded on plagioclase and olivine and enclosing some small plagioclase laths.

In the olivine-rich horizons the augite grains become somewhat larger. At the 786-foot and 1,758-foot levels the crystals attain lengths up to 3 mm.,

and they may hold a few plagioclase laths and some resorbed olivines. In the upper half of the picrite belt the Ca-rich augite forms large poikilitic crystals up to 5 mm. in diameter. They enclose resorbed olivines *but no plagioclase.* The maximum grain size of more than 1 cm. is found in the olivine hyperite just above the picrite. Here some of the enlarged augites enclose plagioclase laths ophitically in addition to some resorbed olivines. Often clusters of the earlier formed augites are still recognisable in such large crystals. Upwards in the quartz hyperite the grain-size diminishes gradually and the augite again forms clusters of small interlocking grains in the lower quartz hypersthene gabbro and the overlying sub-zones. As indicated in Chapter II, the relation between the augite and orthopyroxene changes several times as one proceeds from lower to higher levels within the intrusion. The main change in the crystallisation sequence takes place between the lower and central sub-zones at the 740-foot level. In the picrite the augite is seen to invade the large poikilitic bronzites and in the lower quartz hyperite the augite encloses small prismatic hypersthenes. From the central olivine-bearing sub-zones upwards, however, the orthopyroxene always envelopes augite poikilitically as it invades the augite clusters. Only between the 1,400-foot and 950-foot levels (upper half of the central quartz hypersthene gabbro) the augite seems to enclose and invade the small hypersthene grains, which are probably of an earlier generation. This feature will be discussed later as one of the orthypyroxene-clinopyroxene intergrowths.

The variation in the composition of the augites has been deduced from the axial and extinction angles measured by means of the universal stage. The optic axial angle ranges from $2V_\gamma = 52°$ in the cores of the augites of the picrite to $2V_\gamma = 41°$ for the rims of the augites in the quartz-bearing zones, indicating a decrease in lime content in the less basic phases. This zoning accords with the observations of F. Walker and A. Poldervaart (1949) for the Karroo dolerites in general and with the conclusion of Kuno (1954) that with decreasing temperature, the Ca^{2+} ions are replaced mainly by Fe^{2+} ions.

Feeble normal zoning with decreasing $2V_\gamma$ towards the margin has been noticed in the augites of the Basal Zone. In the Central Zone, axial angles in different parts of a crystal vary slightly and the augites display a patchy extinction. Often a patch in the middle of the grain shows the smallest axial angle of that crystal. In this respect it is noteworthy that reverse zoning has been recorded by H. J. Nel (1940) in the diabase of the Bushveld Complex north of Pretoria.

Only twinned crystals were used for the determination of the extinction angles, and in these the normal to the twinning plane (100) could be determined by matching exactly the interference tints of the twin-halves while rotating about the *b*-crystallographic axis.

The variation in the extinction angle so obtained agrees remarkably well with the variation of the MgO:FeO ratio of the norm (see Diagram I).

Variation in extinction angle is commonly ascribed to variations in the Mg:Fe ratio. Since Winchell's (1951) diagram illustrating the variation in the extinction angles of augites with different Mg:Fe ratios does not agree with the values returned by analysed augites given by H. H. Hess (1949), it was imperative to determine the refractive indices of a few of the augites present in rock types of varying basicity.

Sample		Refractive indices		γ/c
no. 4		$\beta = 1.683 \pm 0.003$		40.5°
no. 10	$\alpha = 1.681 \pm 0.003$	$\beta = 1.692 \pm 0.003$	$\gamma = 1.712 \pm 0.003$	46°
no. 11	$\alpha = 1.681 \pm 0.003$	$\beta = 1.692 \pm 0.003$	$\gamma = 1.714 \pm 0.003$	42°
no. 14	$\alpha = 1.678 \pm 0.003$	$\beta = 1.707 \pm 0.003$		40.5°

F. Walker and A. Poldervaart state that analyses of Karroo pyroxenes indicate that they are all characterised by low sesquioxides. According to the diagram of H. H. Hess (1949) the composition of the augites in the Insizwa profile under consideration vary from about $Wo_{43}En_{47}Fs_{10}$ in the picrite to $Wo_{37}En_{44}Fs_{19}$ in the central quartz hypersthene gabbro.

For augite of this composition the extinction angles of 43° — 45° are too high, according to the diagram of Winchell. It is possible that the confusion may be due to faulty determination of the γ/c values in the past.

(iii) *Orthopyroxene.*

Orthorombic pyroxenes are present in all the rock types studied. They may be distinguished from the augites by their parallel extinction in prismatic sections; their relative freedom from inclusions, slight but distinct pleochroism from pinkish to grey, especially in the prismatic orthopyroxenes of the lower quartz-bearing zones, by their low interference colours, by the lamellar and patchy extinction of the "pseudotwinning lamellae" of sections in the [100] zone (D. L. Scholtz, 1936), the "graphic and lamellar intergrowth" of augitic material in some of them, as ascribed to exsolution by H. H. Hess (1941), A. Poldervaart (1947) and others, and by their poikilitic texture in the rocks from the central olivine zone upwards.

In the chill phase, the orthopyroxene forms very small grains but both the grain-size and the relative proportion of this mineral increase upwards, and, at a height of 16 feet above the basal contact, bronzite, often enveloping augite grains, attains lengths of 1.5 mm. and comprises 23% of the rock. In the overlying picrite, the amount of bronzite, diminishes to about 10% by volume, while it forms large poikilitic plates up to 5 mm. across, which envelop olivine. In the olivine hyperite and quartz hyperite the amount of orthopyroxene increases slightly, while the grain-size diminishes. At the 585-foot level hypersthene reaches a maximum development of 25%. Here the mineral occurs as prismatic crystals 2—3 mm. long, enclosing small plagioclase laths only. Towards the top of these quartz-bearing sub-zones, the prismatic orthopyroxenes become slightly rounded and still smaller (average size 1 mm.).

From the central olivine-bearing zone upwards, however, the orthopyroxene was the last mafic mineral to crystallise, in sub-ophitic and poikilitic networks (3-6 cm. across) forming rims, often very thin, around more or less well-rounded olivines, plagioclase laths and augite grains. The orthopyroxene comprises a fairly constant proportion of about 11% of the rock. The maximum length of 2 cm. across such an optically continuous network was measured in a thin section from the 2,014-foot level, yet the

volume percentage of bronzite was only 9%. F. Walker (1940) found that the orthopyroxenes of the Palisade diabase also displayed this "somewhat unusual habit". These orthopyroxenes are, however, slightly more Mg-rich than the orthopyroxenes of the upper levels of the Insizwa intrusion.

In the olivine-rich horizons at 786 and 1,758 feet the bronzite grains are less crowded with olivine grains, contain smaller plagioclase laths and only occasionally envelop augite grains. In addition, these grains are somewhat smaller, with maximum sizes of 3-3.5 mm.

Some smaller orthopyroxene grains also occur between the 1,200-foot and 1,400-foot levels. But these hypersthenes seem to have crystallised independently of the poikilitic type, and are closely associated with the augite.

The variation in crystallisation sequence of the augite and the orthopyroxenes has been discussed in the preceding paragraph.

The fluctuations in the amount of orthopyroxene seem to depend partly on the amount of olivine present. At the top and bottom of the thick olivine-bearing zones and at the top of the central olivine-bearing zone, the amount of orthopyroxene varies antipathetically with the olivine (see Diagram I). In the olivine-rich horizons at 786 and 1,758 feet an increase in the amount of bronzite has been noticed. Here it is compensated by a decrease in augite and plagioclase. This variation agrees with the concept that the orthopyroxene derived much of its material from the reaction of the early-formed olivines with the remaining liquid (D. L. Scholtz, 1936).

Optic axial angle measurements (see Diagram I) reveal that the composition of the orthopyroxenes not only varies with the proportion of the olivine in the rock, but also with regard to the composition of the olivines and in all probability with that of the augite as well. The most Mg-rich bronzites are those in the middle of the picrite, where the augite is richest in lime and the forsterite content of the olivine is higher than elsewhere. The orthopyroxenes are slightly zonal with $2V_\gamma$ varying from 86° in the core to 81° at the margin. The change in the size of the optical angle of the bronzite away from the centre of the picrite horizon is barely noticeable, but is distinct and abrupt in the vicinity of its margins. Near the basal contact for example $2V_\alpha = 70°$, while in the transitional rocks between the Basal Zone and the Central Zone the optic axial angle ranges from 75° in the cores to 65° in the rims of the orthopyroxenes.

In the central olivine hyperite and the olivine-rich horizon at 1,758 feet, which contain respectively 33% and 26.4% olivine, the optic axial angle of the orthopyroxene ranges from $2V_\alpha = 77°$(core)—70°mantle to $2V_\alpha = 71°$(core)—64°(mantle).

Between these zones there is an increase in the ortho-ferrosilite content with maxima in the olivine-free sub-zones. In the lower olivine free sub-zones the most iron-rich hypersthenes occur near the bottom of this sub-zone, with $2V_\alpha$ ranging from 59° in the cores, to 49° in the rims. Under crossed nicols this strong zoning is immediately recognisable even in unorientated sections because of the increase in the interference tints towards the margins of the crystals.

Pseudo-twinning lamellae.

The orthopyroxenes throughout the intrusion show the lamellar pseudo-twinning structure as described by D. L. Scholtz (1936). This structure,

best observed in sections lying in the [100] zone, appears to be due to the division of the mineral into extremely fine plates, less than 0.005 mm. thick, which are arranged parallel to the plane (100). Under crossed nicols, suitably orientated sections display a fine striation which is due to the alternating extinction positions of successive lamellae. On the basis of optic orientation, two sets of lamellae may be recognised with extinction inclined respectively at plus 3° to plus 6° and minus 3° to minus 6° to the composition face (100) in (010) sections. The extinction position of one of the sets therefore differs by 6° to 12° from that of the other; when best developed the lamellae of the two sets are of identical thickness, neither set tending to favour a particular part of the crystal. In addition, they extend continuously across strongly zoned crystals and terminate abruptly against the crystal margins without diminution in thickness. Some crystals, however, show irregularly distributed patches, comparatively free from lamellae and consist of material corresponding in optic orientation with either one or other of the two sets of lamellae, so that under crossed nicols, the overall extinction of such crystals is patchy.

The material comprising the alternating sets of lamellae is *identical* in refractive indices and optic axial angle and shows similar interference tints. It must be emphasised then *that orthopyroxenes showing this structure are homogeneous,* and, but for orientation, no differences in the physical and optical properties of the material comprising this structure can be detected.

Fig. 2. Approximate (001) section of orthopyroxene showing traces of prismatic cleavage and pseudo-twinning lamellation. Crossed nicols, × 200 (basal hyperite, Insizwa).

Although a certain amount of confusion still exists in the literature, it is clear that the pseudo-twinning structure differs markedly from that resulting from the intergrowth of monoclinic pyroxene with orthopyroxene, such as has been described in the orthopyroxenes of the Stillwater Complex by H. H. Hess and A. H. Phillips (1938). This intergrowth is due to the presence of very fine platelets of clinopyroxene in the form of closely-spaced sheets, 0.002 mm. thick and 0.01 mm. apart. The optic plane of the clinopyroxene inclusions is perpendicular to that of the orthopyroxene host, while the platelets show higher birefringence and large extinction angles of up to 40°. H. H. Hess and A. H. Phillips regard this intergrowth as a result of the exsolution of augite from the orthopyroxene, a relation so obvious that is was never open to doubt. But failure to recognise the fundamental difference between the exsolution platelets and the pseudo-twinning lamellation induced them to conclude erroneously that the two structures were identical.

This statement can easily be verified without employing the universal stage. According to H. H. Hess and A. H. Phillips (p. 452) no lamellae are visible in (001) sections of orthopyroxenes characterised by the pseudo-twinning lamellation. This, however, is not so for sections slightly inclined to the basal plane (see Fig. 2).

As a matter of fact the lamellae are most distinct and as sharply defined as in all other sections lying in the [100] zone,* that is perpendicular to the optic axial plane, the only difference pertaining to the size of the extinction angle.

An examination on the universal stage of several thin sections of the Stillwater Complex from the collection at this University shows that the pseudo-twinning lamellae are also present in the orthopyroxenes. However, fine exsolution platelets of clinopyroxene orientated parallel to the pseudo-twinning plane (100) are also frequently present. Because they differ appreciably in optical properties from the host, their distinct, sharper and more obvious structure tends to obscure the feebler and less apparent pseudo-twinning lamellation.

A similar relationship between augitic exsolution platelets and orthopyroxene lamellae has been observed in the rocks from certain sub-zones at Insizwa.

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* H. H. Hess and A. H. Phillips also recommend the adoption of the orientation proposed by Groth, Goldschmidt, Rosenbusch and others rather than that proposed by Hintze, Dana and Winchell. This suggestion, already made by numerous mineralogists and petrologists, would if adopted, serve to eliminate much of the existing confusion in the literature of the pyroxenes. The axial ratios listed below, indicate that the last named authors adopt one practice in the case of clinoenstatite and another for enstatite. It will be seen that the intercept on the *a*-crystallographic axis of the former mineral is larger than that used on the *b*-crystallographic axis, while the reverse relation holds good in the case of enstatite. This procedure not only interchanges the position of the *a* and *b* axes, but also tends to create confusion and conceal important morphological and leptological similarities.

Mineral	*Axial ratio*	*Authorities*
Enstatite	1.0308 : 1 : 0.5885	Groth, Rosenbusch, Goldschmidt, Iddings, Johannsen, Niggli and others.
	0.9702 : 1 : 0.5710	Hintze, Dana, Winchell.
Clinoenstatite	1.033 : 1 : 0.591	Rosenbusch, Winchell.

The twinned relationship between the lamellae of this structure is best revealed by a stereographic plot of the optic and crystallographic directions as determined from investigation on the universal stage. Such a diagram shows that the α-aether axes always coincide, but that the -γ and β-aether axes of the two sets are inclined to one another at angles varying from 6° to 12°. Further, the *c*-crystallographic axis constructed from the great circle of the prismatic cleavage normals, or from β and one or both (110) normals, is usually found to fall midway between the γ-aether axes of alternating sets of lamellae. The relationship may thus be explained as a type of polysynthetic Carlsbad twinning about the *c*-crystallographic axis or hemitropism about the normal to (100), the composition face being parallel to (100). This of course suggests that these pyroxenes are not strictly orthorhombic, but that owing to extreme multiple twinning, the crystals as a whole, display an apparent orthorhombic character. It is further quite conceivable, as has been indicated by D. L. Scholtz (1936) and others, that if the lamellae were of submicroscopic dimensions the true monoclinic character of the mineral would be effectively concealed.

The careful and detailed X-ray investigation of such lamellar orthopyroxenes enabled N. H. Henry (1942) to prove conclusively, that the lamellation is not due to the presence of minute exsolution platelets or films of augite or diopside, as postulated by H. H. Hess and A. H. Phillips. He ascribes the origin of the structure to translation parallel to [001] on (100), accompanied by bending about [010], that is, a type of "Biegegleitung". N. H. Henry also holds that the lamellation owes its origin to deformation during the crystallisation of the orthopyroxenes.

C. M. Schwellnus (1939) who investigated more than 1,500 grains of bronzite and hypersthene ($2V_\alpha = 90°\text{-}55°$) in the Bushveld Complex by means of the universal stage, states that all the pyroxenes of the Basal, Critical and Main Zones of the intrusive are characterised by lamellar structures which are not to be confused with exsolution platelets of diopside pyroxene usually present in small quantities. The occurrence of bent crystals of hypersthenes in a number of localities points to the deformation of the rocks containing them, but the observation that some crystals of orthopyroxene are characterised by lamellar cores and non-lamellar mantles and vice versa leads the writer to believe that the characteristic lamellation of the orthopyroxene could not have originated by translation or "Biegegleitung".

In discussing the pseudo-twinning lamellae of the orthopyroxenes N. L. Bowen and J. F. Schairer (1935) state that "the inclined extinction as measured against the twinning is not sufficient to disprove orthorhombic symmetry since twinning on the dome (014) gives such an effect".

According to H. Rosenbusch, A. Johannsen, P. Niggli and K. Chudoba the lamellar structure in question is parallel to (100). D. L. Scholtz deduced the crystallographic orientation of the lamellae and optic axial plane by the careful universal stage measurement of the prismatic cleavage angles and the relative positions of the aether axes α which incidentally is the normal to (010). As already pointed out, H. H. Hess and A. H. Phillips questioned the accuracy of the first determination but agreed afterwards with him that the axial plane as well as the lamellar structure is parallel to (100). Subsequent, independent work by C. M. Schwellnus, H. J. Nel, the writer and others, served to confirm the conclusion of D. L. Scholtz.

N. L. Bowen and J. F. Schairer believe that the lamellation is a form of polysynthetic twinning but not on the (100) face. That the lamellae in question cannot possibly be due to twinning on the dome (014) or rather (104)* as suggested by N. L. Bowen and J. F. Schairer is only too clear. Since the faces of this form are very nearly parallel to the basal pinacoid it is to be expected that the lamellation should be ill-defined or invisible on sections cut nearly parallel to (001). In practice, however, the reverse is true (see Fig. 2).

If the pseudo-lammellar structure is not due to exsolution of clino-pyroxene or translation it can only be regarded as a multiple twinning phenomenon. The evidence in favour of this conclusion can be briefly summarised as follows:

1) The alternating lamellae which characterise the orthopyroxenes of several plutonic intrusions are identical in composition, optical properties and leptological structure.
2) The approximate symmetrical extinction displayed by adjacent lamellae about [001] or $\perp$ (100) as well as the fact that [010] is always common to both, is characteristic of twinned individuals of monoclinic symmetry.
3) The experiments of N. L. Bowen and J. F. Schairer indicate that pseudo-twinned and untwinned natural orthopyroxenes are characterised by identical thermal behaviour.
4) Natural orthopyroxenes have been transformed into true clinopyroxenes in the laboratory. The recent work of R. W. Goranson (1938) and N. L. Bowen and O. F. Tuttle (1950) clearly indicates that the crystallisation equilibria of dry alkali feldspar melts are practically unchanged except that they function at very much lower temperatures in a wet pressure system. If the sluggish transformation of bronzite in a dry system is greatly facilitated by the addition of a flux like NaF or LiF, it may reasonably be expected that the high clinopyroxene-orthopyroxene transformation temperatures may also be lowered within the range of magmatic temperatures, in wet or volatile enriched magmatic systems.
5) Clinoenstatite and clinohypersthene are almost invariably polysynthetically twinned on (100). In orthopyroxene the finer pseudo-twin lamellation is characteristic and also parallel to (100).
6) According to B. E. Warren and D. I. Modell (1930) the unit cell of orthorhombic enstatite may be regarded as being composed of two opposed monoclinic cells in which the like and mutual *bc* plane is also parallel to (100).
7) The concept of polysymmetry in the case of enstatite series has not been disproved. The interruption of the alternating succession of monoclinic sub-cells in enstatite by the introduction of clinoenstatite unit cells according to the L. Atlas (1951) pattern would accentuate the monoclinic element in the enstatite, thereby initiating the lamellar structure of the orthopyroxenes as well as accounting for the fact that the extinction angles of the polysynthetic twinning lamellae are always less than the extinction angle of pure clinoenstatite.

* (104) according to the preferred orientation of Goldschmidt and Groth.

(iv) *Relation to olivine.*

The fact that over a wide range of composition of basaltic magmas olivine crystallises out, but is later partly or completely redissolved by reacting with the liquid to give lime-poor pyroxene, is well known from field observations, and from the investigations of the binary system $MgO-SiO_2$ by N. L. Bowen and O. Anderson (1944).

This reaction clearly accounts for the poikilitic habit of the pyroxenes in the Basal Zone and possibly also for the pseudo-poikilitic habit of the orthopyroxene in the central and upper sub-zones and some horizons of the lower sub-zones.

Support for this conclusion is found in the relation between the compositions of the olivines and both the pyroxenes occurring in the same rocks. To compare the compositions of the coexisting olivines, augites and orthopyroxenes, the chemical compositions deduced from the optical properties of these minerals have been plotted in Figure 3. This diagram shows that in the Basal Zone (samples 2-5) both the pyroxenes are more iron-rich than the olivines of the same horizons, and that in the Central Zone (samples 6-17) the augites seem to be more magnesium-rich than the coexisting olivines, but the orthopyroxenes are of an intermediate composition. The latter fact supports the assumption that the orthopyroxenes grew at the cost of olivines and augite.

The fact that olivines may occur together with more magnesian-rich pyroxene has been recorded before. The experimental work of N. L. Bowen and J. F. Schairer (1935) shows that in a $MgO-FeO-SiO_2$ melt all olivines, except forsterite, are in equilibrium with more magnesia-rich, lime-free pyroxenes. H. Ramberg and G. de Vore (1951) compare those results with the relation between the composition of natural co-existing olivines and orthopyroxenes in rocks. They found that in the magnesia-rich assemblages the olivine is usually more magnesia-rich than the orthopyroxene, but that in the more iron-rich assemblages, those which hold more than 35% iron-silicates in their average composition, the reverse relation obtains at an increasing rate. H. Ramberg and G. de Vore ascribe this to a mutual interchange of magnesia and iron between the two phases when they cool in the absence of free silica.

The small composition range of the augites in comparison with the ranges of the olivines and orthopyroxenes is striking. According to H. H. Hess (1941) the same relation obtains in the augites and orthopyroxenes of other basic rock suites. He draws attention to the fact that the joins representing the composition of augites and associated orthopyroxenes plotted in an En-Fs-Wo-diagram, seem to intersect the En-Wo-join near $En_{25}Wo_{75}$. He failed to account for this relation. In the case of the Insizwa pyroxenes under consideration a similar relation appears to exist (see Fig. 3(a)).

As is mentioned by F. Walker and A. Poldervaart (1941) for the Karroo dolerites in general, the crystallisation period of orthopyroxene and lime-poor pigeonite overlaps that of augite. In the Insizwa profile under consideration augite envelopes the orthopyroxene cores in the lower sub-zones while in the central and upper sub-zones the orthopyroxene is always preceeded by augite. The crystallisation sequence, however, appears not to depend on the composition of the orthopyroxene only, as the augite-enveloped orthopyroxene in the lower

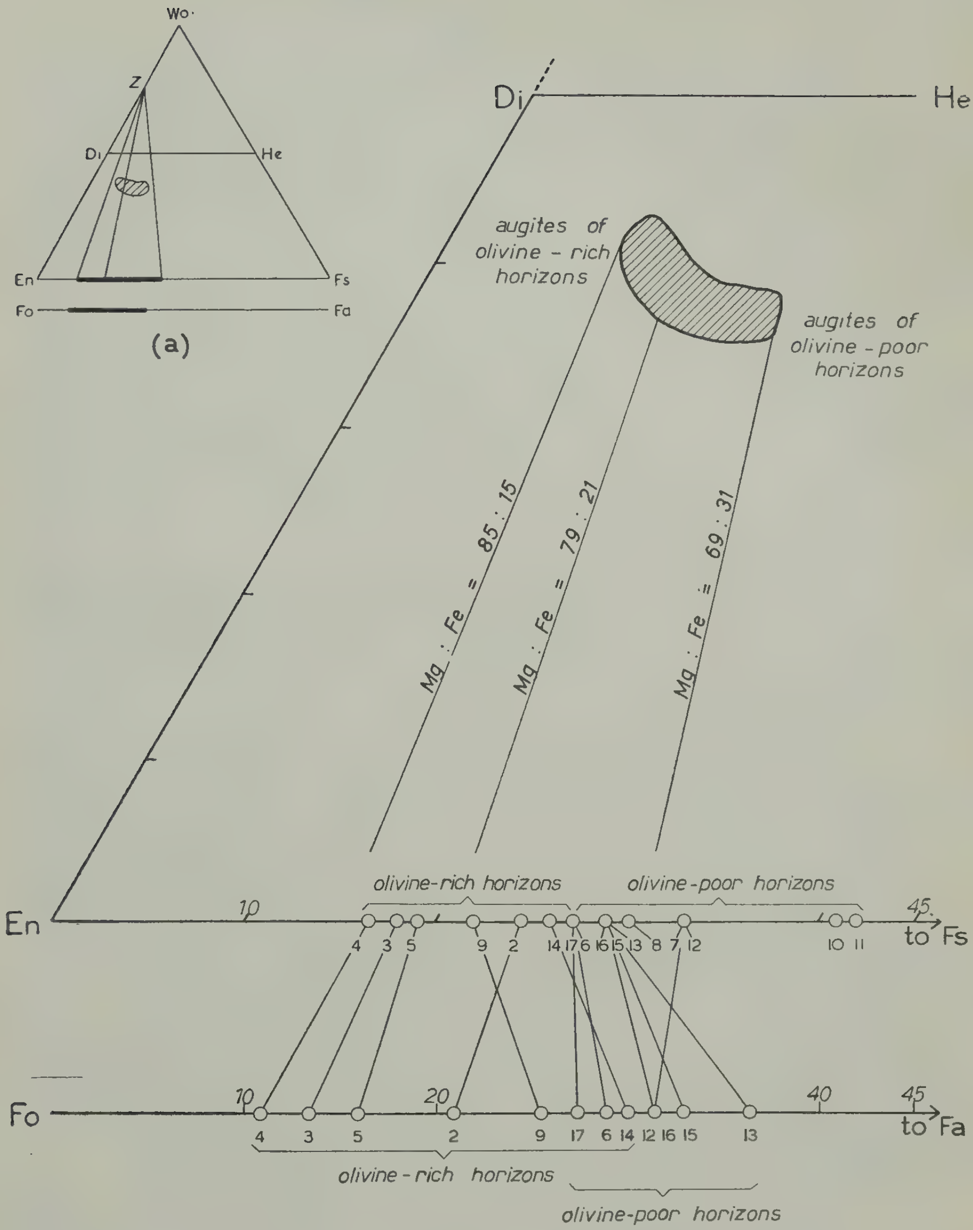
Wo
Z
Di
He
En
Fs
Fo
Fa
(a)
Di
He
augites of olivine - rich horizons
augites of olivine - poor horizons
Mg : Fe = 85 : 15
Mg : Fe = 79 : 21
Mg : Fe = 69 : 31
olivine-rich horizons
olivine-poor horizons
En
10
45
to Fs
4 3 5 9 2 14 17 6 16 15 13 8 7 12 10 11
Fo
10
20
40
45
to Fa
4 3 5 2 9 17 6 14 12 16 15 13
olivine - rich horizons
olivine-poor horizons

Fig. 3

sub-zones has compositions as iron-rich as Fs_{30}, while in the central olivine hyperite the enveloping orthopyroxene has a composition of Fs_{23}.

(v) *Orthopyroxene-clinopyroxene intergrowths.*

The crystallisation of the lime-poor metasilicates has been intensively studied over the past 20 years. The fundamental laboratory investigation of the ternary $MgO\text{-}FeO\text{-}SiO_2$ system by N. L. Bowen and J. F. Schairer in 1935, has been largely substantiated by field observations. Since 1938, the most important discussions of the natural occurrences, with special reference to the inversion of pigeonite to orthopyroxene and the orthopyroxene-clinopyroxene intergrowths, have been made by H. H. Hess, A. H. Phillips, F. Walker, A. B. Edwards, D. Guimaras, A. Poldervaart, L. Atlas, H. Kuno and K. Nagashima, and they arrived at the following conclusions:

In appropriate melts, a clinopyroxene phase crystallises initially and inverts to orthopyroxene on slow cooling. The inversion temperature established by N. L. Bowen and J. F. Schairer falls with increase of $FeSiO_3$ content from approximately 1,140° to 955° C. In natural magmas the crystallisation of pyroxene should occur at lower temperatures as a result of the presence of additional components and volatile substances, while the effect of pressure is by no means negligible. Under the circumstances it is believed that magnesian-pyroxenes will crystallise as orthopyroxenes, while the more iron-rich varieties of this series will form pigeonite.

It is stated that the inversion of pigeonite to orthopyroxene takes place "by doubling of the monoclinic cell along the a-crystallographic axis with a slight shift of the SiO_4 chains". This would involve a slight contraction so that the Ca^{2+} ions can enter this structure only up to 3½% of the total $Ca^{2+}\text{-}Mg^{2+}\text{-}Fe^{2+}$ ions or approximately one third of the number present in pigeonite.

It would seem as if the more magnesian pyroxenes displaying evidence of inversion, correspond approximately in composition to $En_{70}Fs_{30}$. H. Kuno (1952) showed that orthopyroxenes of this composition which were in equilibrium with pigeonite across the inversion interval might hold up to 5% Ca^{2+}, while 8% of that element was present in the associated pigeonite.

In 1954 H. Kuno found that the temperature at the time of crystallisation is a more important controlling factor than the Mg:Fe ratio in determining the amount of calcium entering into the pyroxene structure. He asserts that it is primarily the substitution of Si^{4+} by Fe^{3+} in the Si-O chains which renders it possible for Ca^{2+}, Mg^{2+} and Fe^{2+} to enter the pyroxene structure in all proportions. In this connection it is interesting to note that N. L. Bowen and J. F. Schairer showed that the amount of Fe_2O_3 present in the pyroxene region of the $MgO\text{-}FeO\text{-}SiO_2$ system, increases with the FeO content.

The "intergrowths" of clinopyroxene in hypersthene are ascribed to the segregation of diopsidic or augitic material. This material may have been expelled;

(i) from orthopyroxene which held the excess of $CaSiO_3$ in solid solution at high temperature but exsolved it upon slow cooling, or

(ii) as a direct result of the inversion of pigeonite to hypersthene, or

(iii) from pigeonite before inversion took place. (Diopsidic sheet-like segregates in pigeonite were observed by A. M. J. de Swart and L. J. Murray (1944), as well as by J. S. van Zyl (1950). Likewise H. Kunō (1951) found "pigeonites with admixtures of augite").

These relations are portrayed in the diagram of A. Poldervaart and H. H. Hess.

Exsolution is believed to take place only upon slow cooling, the clinopyroxene being expelled along definite planes within the orthopyroxene. A. Poldervaart and H. H. Hess believe that the relation between the orientation and thickness of the exsolution platelets and the crystal structure of the different pyroxene hosts can be explained by assuming that unmixing will occur, so that the exsolution products and the host have common structural planes, while the orientation and thickness of the lamellae are determined by the ease of access of the ions to the exsolution plane. According to this view the exsolution of clinopyroxene from orthopyroxene will therefore take place so that the *b*- and *c*-crystallographic axes of the ortho- and clinopyroxene are parallel, while the clinopyroxene forms thin lamellae parallel to their common (100) plane. The exsolution plane of augite in pigeonite is normally (001) and the platelets are coarser than those appearing along the (100) plane in orthopyroxene, since the migration of ions takes place respectively parallel to and across the SiO_4 chains.

During the inversion of pigeonite, the orthopyroxene will tend to develop in such a manner that its *b*- and *c*-crystallographic axes are respectively parallel to those of its host. But in exceptional cases the *c*-axis of the orthopyroxene may be inclined at 45° to that of the pigeonite. After the inversion the relic (001) plane is parallel to a hypothetical dome plane of the orthopyroxene, nearly but not exactly parallel to (101), according to A. Poldervaart and H. H. Hess, but the writer believes that the symbol (102)* is more appropriate. The platelets in the orthopyroxenes of hypabyssal rocks will be less well developed than those in the more slowly cooled plutonic equivalents and trails of irregular platy blebs parallel to the relic (001) plane instead of the regular platelets may occur in the hypersthene.

For the sake of convenience the various ortho- and clinopyroxene associations observed in the Insizwa intrusion may be briefly described with reference to Figure 4 in order to accentuate the salient characteristics and differences with special emphasis on the principal types of intergrowths.

(a) An aggregate in which a single highly irregular and lamellar pyroxene displaying traces of the (110) cleavage is seen to enclose anhedral and embayed grains of augite, some of which are in optical continuity. The structurally continuous grains may be recognised in Figure 4(a) by the parallelism of the fine ruling, which represents the trace of the (001) plane of the clinopyroxene. A photograph illustrating this type of aggregate is represented in Plate I, pm. 4.

(b) The aggregate consists of a single irregular and lamellar orthopyroxene crystal enclosing differently orientated anhedral to embayed grains of augite and olivine. Well developed films of augite are orientated parallel to (100) and the pseudo-twin lamellation of the host. See Figure 4(b) and Plate I, pm. 11.

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* Based on Goldschmidt's value for the intercept on the *a*-crystallographic axis.

Fig. 4

(c) An aggregate, as in Figure 4(b), but in which one of the included grains of augite is in optical continuity with the two sets of intergrowths in the orthopyroxene host. The additional set, not previously described and consisting of well developed platelets or platy blebs 0.002-0.008 mm. thick and 0.02-0.04 mm. apart, is tabular and parallel to (001) of the master augite grain and approximately parallel to the (102) plane of the orthopyroxene. If the included master crystal of augite is twinned, its composition face (100) is parallel to (100) of the orthopyroxene, and a herring-bone pattern of domal platelets is developed in the host. The (100) films appear to attain their best development in those parts of the orthopyroxene crystals which are devoid of platelets.

(d_1) In its simplest form this association comprises a single small highly irregular and embayed grain of orthopyroxene partially or completely enveloped by a single crystal of augite with the *b*- and *c*-crystallographic axes in common. Well developed platelets of clinopyroxene, flattened parallel to (001) and of the same orientation as the augite, appear in the orthopyroxene. They very nearly follow the (102) dome plane of the host, which also appears to exploit the augite along its (001) plane, but it is noteworthy that films of pyroxene in optical continuity with the orthopyroxene may also appear in and parallel to the (001) plane of the augite (see Fig. 4(d)).

(d_2) This association may be regarded as a cluster of several differently orientated, but less common, single d_1 binary units.

(e) An aggregate consisting of a large irregular and lamellar crystal in optical continuity with several small interstitial and manifestly detached grains of hypersthene partially enveloped by a large anhedral crystal of augite simply twinned on (100) with its composition face parallel to (100) of the orthopyroxene. Basal platelets and films, both in optical continuity with the augite crystal, tend to follow respectively the domal planes and the twin lamellation not only in the host crystal, but also in some of the manifestly detached grains of the orthopyroxene. Films and platelets bear the same relation to one another as in 4(c), while the mutual inclination of the two sets of augite platelets in the host also gives rise to the typical herring-bone structure.

(f_1) In this type of aggregate a single lamellar crystal of orthopyroxene is intimately associated with numerous diversely orientated grains

Fig. 4. Illustrating some of the different types of orthopyroxene-clinopyroxene aggregates in the Insizwa intrusion. In all the figures the c-crystallographic axes of the orthopyroxenes are vertical while clinopyroxenes may be distinguished by closely spaced ruling parallel to their basal planes (for explanation see text).

of augite. In Figure 4(f_1) four such crystals of augite are represented, in which the finely ruled lines not only serve to reveal the traces of their (001) planes but also their haphazard orientation. Four sets of arbitrarily orientated platelets, tabular parallel to (001) and each set in optical continuity with one of the four augite grains, now appears in the orthopyroxene host.

(f_2) An association in which a single anhedral and embayed crystal of orthopyroxene is partially or completely enveloped by two differently orientated crystals of augite. The films throughout the orthopyroxene, and the platelets in optical continuity with the master augite grain A, bear the usual relation to their host, whose *b*- and *c*-crystallographic axes coincide with those of the crystal A. The other grain of augite, as well as its related set of platelets, displays an entire lack of orientation with respect to the structure of the orthopyroxene crystal. It should be noted that no film structure related to this augite is present in the host-crystal, but the augite films that are present are all related to the structure of the master augite A.

(f_3) In aggregates of the type represented by this figure, a single large and highly irregular crystal of hypersthene characterised by the presence of numerous arbitrarily orientated sets of platelets and platy blebs of different inclination scattered throughout the substance of the orthopyroxene. Augite inclusions may or may not be structurally related to any one of these sets of intergrowths. In Fig. 4(f_3) neither the twin lamellation of the hypersthene nor films of clinopyroxene are visible since the section is cut parallel to (100). This conclusion is also substantiated by the fact that the hypersthene shows the highest interference tints characteristic of sections parallel to the optic plane.

From the foregoing description of the different kinds of pyroxene aggregates investigated on the universal stage and from the petrography of the ferromagnesian minerals present in the Waterfall Gorge section of the intrusion, the following general conclusions may be drawn:

1. When the amount of orthopyroxene in the rock exceeds that of clinopyroxene, it appears to have crystallised first and vice versa. This observation, however, does not apply to the upper olivine hyperite.
2. The range of crystallisation of the orthopyroxene in the rocks under consideration overlaps that of the clinopyroxene.
3. Orthopyroxene is always characterised by the pseudo-twin lamellar structure parallel to (100).
4. Orthopyroxene tends to occur in larger single crystals than clinopyroxene when its concentration in the rock is lower than that of clinopyroxene.
5. When orthopyroxene occurs in large crystals it almost invariably holds inclusions of clinopyroxene, plagioclase and olivine if present, in pseudo-poikilitic fashion.
6. Three principal types of clinopyroxene-orthopyroxene intergrowths occur in the rocks under consideration and they may be classified according to their crystallographic orientation in the orthopyroxene host as well as according to their relative dimensions. We can distinguish:

(a) attenuated platelets of films orientated parallel to the (100) plane of the orthopyroxene.

(b) platelets or platy blebs orientated almost parallel to the (102) plane of the orthopyroxene.

(c) platelets or platy blebs of random orientation.

7. The intergrowths may be present, but are often absent in those hypersthenes or rims of strongly zonal orthopyroxenes with compositions ranging from Fs_{33} to Fs_{50}. Generally speaking the abundance and the randomness of the intergrowths increase with increasing acidity of the rocks.

8. Clinopyroxene films are often present in orthopyroxenes of the composition range Fs_{33} to Fs_{50}. The films follow no other direction than (100) in the orthopyroxene host irrespective of the number of associated or included grains of clinopyroxene. Under the most favourable circumstances the films will be found to be in optical continuity with a master clinopyroxene grain whose *b*- and *c*-crystallographic axes are parallel to those of the host.

9. The tapering nature and the relation of the films of clinopyroxene to the orthopyroxene are such as to suggest that they represent true exsolution products of the host while in the solid state.

10. Orthopyroxenes often selectively exploit the margins of clinopyroxene crystals along their basal planes, irrespective of the orientation of the grains.

11. Although the orthopyroxene often exploits the basal directions of augite, the relic platelets parallel to the basal directions are rarely in contact with these augites, save in the case of the binary units (Fig. 3(d)).

12. Thin films of orthopyroxene in optical continuity with the main crystal may occasionally be observed along the (001) planes of clinopyroxene.

13. The clinopyroxene grains of the rocks of the central and upper subzones appear to exhibit different stages of replacement by orthopyroxene.

14. The relative abundance of clinopyroxene and orthopyroxene in the rock appears to exercise a certain measure of control over the nature of the aggregate.

    If on the one hand augite is volumetrically more abundant than hypersthene, isolated binary units may appear which tend to form arbitrarily orientated aggregates of such crystal pairs (see Fig. $4(d_2)$), while large crystals of orthopyroxene characterised by one or numerous diversely orientated sets of clinopyroxene platelets, are present in all rocks which exhibit "intergrowths".

    On the other hand, if orthopyroxene is more abundant than clinopyroxene, and intergrowths do occur, then the intergrowths, either with the normal or with a random orientation, are confined to the rims of the orthopyroxene crystals. These rims have compositions ranging from Fs_{35} to Fs_{42} in the section under consideration.

15. Platy blebs or platelets of clinopyroxene in orthopyroxene tend to be tabular parallel to (001).

16. The platelets of clinopyroxene may or may not be orientated approximately parallel to the (102) domal plane of the orthopyroxene crystal in which they appear.
17. When a single large irregular crystal of orthopyroxene is associated with a number of differently orientated grains of clinopyroxene, only one of them, whether a single crystal or a twin, can function as a master crystal whose *b*- and *c*-axes coincide with those of the orthopyroxene.
18. Sets of platelets or platy blebs of clinopyroxene in optical continuity with a master crystal of augite are always aligned approximately parallel to the (102) plane of the orthopyroxene host. If the master augite is twinned, two sets of platelets parallel to the (001) plane of the two halves of the twin appear in the orthopyroxene constituting the well known and frequently developed herring-bone structure.
19. A single large irregular and interstitial crystal of orthopyroxene may be characterised by the presence of up to seven or even more sets of clinopyroxene platelets. There is only one master set parallel to the (102) domal plane of the host but it need not necessarily be visible in every section. Grains of clinopyroxene are less frequently associated with orthopyroxene crystals of this type.
20. The pseudo-poikilitic habit of the orthopyroxene, though partly concealing the relation between the exsolution platelets and films and the parent pyroxenes from which they originated, nevertheless appears to be useful in establishing the relative ages of the exsolution products, inversion phenomena and the reaction which resulted in the production of the poikilitic texture of the orthopyroxene.

This brief summary of the mutual relations of the orthopyroxenes and clinopyroxenes and the various types of intergrowths between them clearly indicates that the current and generally accepted views of H. H. Hess and others cannot be applied without reserve. The writer feels that undue stress has been laid on inversion phenomena, primarily at the expense of replacement processes, which in some instances appear to afford the only reasonable explanation of the origin of the complex system of intergrowths observed in single crystals of orthopyroxene.

Further detailed investigations now in progress along other surveyed profiles of this fascinating intrusion will no doubt serve to cast more light on this rather confused aspect of petrography.

Accessory minerals.

The minor mineral constituents of the Insizwa rocks are apatite, magnetite, ilmenite, sulphides, biotite, hornblende, alkali feldspar and quartz. Their volumetric modes, normative values and distribution in the section of the intrusion under consideration are given in Diagram I.

Magnetite and apatite appear to have been the first minor constituents to crystallise. Small magnetite octahedra are enclosed in all the principal minerals. Apatite has been detected only in the late interstitial minerals, such as alkali feldspar and quartz and the outer zones of plagioclase laths.

Irregular ilmenite rods and tiny sulphide pellets are usually associated with biotite, alkali feldspar and quartz.

Apatite occurs throughout the intrusion in minute quantities usually less than 0.1%. In the Basal Zone it appears as thin needles, less than 0.02 mm. thick, but in the Central Zone it crystallises as stouter prisms, 0.1 mm. thick and up to 3-5 mm. long. They are sometimes anhedral. Crystals of apatite are largest and most abundant in the lower quartz hyperite. Here it is sparsely distributed in relatively coarse grains up to 0.25 mm. in diameter, but comprises only 0.3% by volume of the rock.

Magnetite usually occurs as small octahedra and rarely in the form of irregular grains together with ilmenite. It is most abundant in the olivine-rich rock types and seldom occurs in the olivine-free sub-zones of the intrusion. Magnetite is seldom associated with biotite.

Ilmenite seems to be more abundant in the Central Zone than in the Basal Zone. It is usually enveloped by aggregates of biotite.

The sulphides, often together with ilmenite, occur as irregular tiny pellets associated with the late interstitial minerals, especially biotite.

In polished sections the sulphides are seen to consist mainly of pyrrhotite with α and β components and flames of exsolved secondary pentlandite. Small amounts of primary pentlandite and chalcopyrite are intimately associated with the spangles of pyrrhotite. The chalcopyrite is twinned, and the direction of the twin lamellae appears to control the orientation of the flakes of valleriite which are occasionally present.

The sulphides seem to be concentrated not so much in the picrite itself as below and on top of it and in the central olivine hyperite. The norms record 1.67% and 0.85% Fe_7S_8 in samples 2 and 9 respectively. Available assay results indicate that a disseminated sulphide zone 240 inches thick and a few feet above the basal contact holds 0.2% of metallic nickel. Another mineralised zone on the 370-foot level returned 0.1% Ni and 0.16% Cu over 24 inches.

Biotite occurs as strongly pleochroic red-brown flakes closely associated with the sulphides and ilmenite. This association is also demonstrated by the general sympathetic relation between the modal ore and biotite curves of the Central Zone, while in the picrite, where the bulk of the relatively large amount of ore is magnetite, the biotite is present only in small quantities.

The intimate association of red-brown biotite and ilmenite, the consistent higher concentration of normative ilmenite over total modal ore, the absence of any other titaniferous minerals in the rocks studied, and the general sympathetic character of the modal biotite and normative ilmenite curves (see Diagram III) appear to corroborate A. J. Hall's (1941) conclusion that red-brown biotites are titaniferous. F. Walker and A. Poldervaart come to the conclusion that the presence of chromic oxide is responsible for the red-brown colour of biotite in the Karroo dolerites. They reason that the biotite is derived from the magnesian olivines and orthopyroxenes and would therefore also be rich in magnesia, the presence of which would mask the colour effect caused by titania. At the end of his publication A. J. Hall, however, points out that a rather high magnesium content will not mask the red colour of biotite as long as the titanium content is high and provided the iron-content is low. In this connection it should be pointed out that since the biotite is late in the crystallisation sequence, it need not necessarily be

as rich in magnesia as F. Walker and A. Poldervaart assume it is. Further work is necessary to settle this point.

The paragenesis of biotite and the sulphides is rather complicated. The mineral usually envelopes the ore, but the ore penetrates and exploits the cleavage planes of the biotite. Therefore it seems that the ore was separated at a relatively early stage but only crystallised after the last of the silicates, and reacted with them before congealing.

In the chill phase, biotite, though abundant, does not appear to be associated with ore.

Hornblende occurs in small quantities, associated with the pyroxenes and biotite, mainly in the rocks devoid of olivine. Only the lower quartz hyperite holds more than 0.5% by volume of this mineral. Hornblende is feebly pleochroic in bluish and brownish green. It has an optic axial angle $2V_\alpha = 70°\text{-}80°$ and has an extinction angle $\gamma/c = 16°$. The *b*-axis of the hornblende grains and the associated augite coincide, while those associated with the orthopyroxene have the [001] and [010] directions in common.

Uralite also occurs in minute quantities in the olivine-bearing rocks in which serpentine veins cross the pyroxenes and biotite.

Alkali feldspar is present in minute quantities throughout the intrusion as interstitial grains with an optic axial angle $2V_\alpha = 30°\text{-}50°$. It attains its greatest concentration in the lower half of the Central Zone with maxima in the quartz-bearing sub-zones. The interstitial distribution within the rocks is indicative of the late crystallisation of the alkali feldspar.

Quartz is also interstitial and usually graphically intergrown with alkali feldspar. Isolated anhedral quartz grains were only observed in the chill phase.

IV. CHEMISTRY

It is remarkable that in spite of the numerous publications on, and world-wide interest in, the differentiated intrusives of East Griqualand and Pondoland, comparatively few chemical analyses of the most important rock types are available. Up to 1936 only seven largely inferior silicate analyses were available. In that year D. L. Scholtz published the results of eight additional superior chemical analyses of rock samples, mainly from the Basal Zone of the Insizwa intrusion, the investigation of which constituted the bulk of the substance of his memoir.

Scarcity of outcrops and the heavy overburden along the steep and precipitous flanks of the Insizwa Range probably constituted the principal obstacle to the petrological investigation of the lower part of the main gabbroic body of the intrusion until the EX-core of the borehole W.G. 5 situated below the cliffs on the right bank of the Waterfall Gorge streamlet was made available for study through the courtesy of Dr. L. T. Nel, former director of the Geological Survey of the Union of South Africa.

After the initial petrographic investigation of the entire rock suite, nine samples from the core, and eight of the thirty-five samples collected along the surveyed profile in the field (see Fig. 1), were submitted for analysis to the Division of Chemical Services, Pretoria. This work was undertaken by Mr. A. Kruger and the values returned by the analyses are listed in Table IIa.

The doleritic and basaltic magma of the Karroo area is generally, and probably correctly, regarded as typically tholeiitic in character, although the similarity in chemical composition has been somewhat overstressed in the past. That precise agreement is unlikely appears to be a natural phenomenon if the plateau basalt magma type is genetically regarded as a sial-contaminated derivative of the primary olivine basaltic magma.

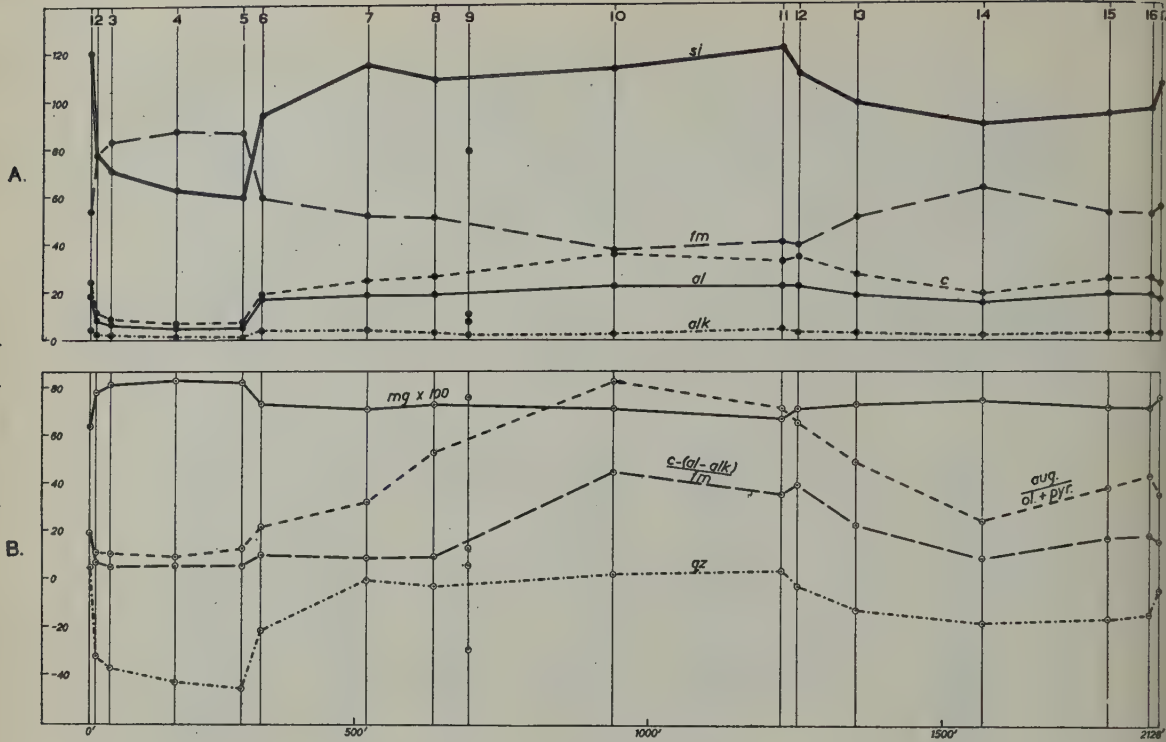
The close agreement between F. Walker's (1949) average analysis of Karroo dolerite and the average dolerite in the immediate vicinity of the Insizwa (Table 1, C and D) corroborates D. L. Scholtz's contention that the Insizwa lopolith originated from the passive intrusion of undifferentiated and uncontaminated magma.

Table I.

	A	B	C	D	E
SiO_2	52.6	52.9	52.3	52.5	51.6
Al_2O_3	13.0	13.8	15.6	15.4	15.0
Fe_2O_3	1.9	1.8	0.4	1.2	3.4
FeO	7.0	8.3	9.6	9.3	9.6
MgO	13.4	10.0	6.7	7.1	4.9
CaO	7.7	9.9	10.3	10.3	8.9
Na_2O	2.6	1.4	2.3	2.1	2.4
K_2O	0.8	0.6	0.9	0.8	1.1
TiO_2	0.6	0.9	1.2	1.0	2.7
P_2O_5	0.1	0.2	0.2	0.1	0.2
MnO	0.3	0.2	0.5	0.2	0.2
	100.0	100.0	100.0	100.0	100.0

A. Chill phase, Morley's Adit, Waterfall Gorge	W. Goodchild
B. Chill phase, base of B.H. W.G. 5, Waterfall Gorge	A. Kruger
C. Average Karroo Dolerite, Insizwa area	E. Radley
D. Average Karroo Dolerite (43 analyses)	F. Walker
E. Average Tholeiite (32 analyses)	F. Walker

A comparison of the chemical composition of the chill phase at the base of the intrusion with that of the average Karroo dolerite indicates that it is distinctly more magnesian, but less aluminous and alkaline.
That this is a real characteristic of the chill phase is verified by the old analysis of similar material collected by W. H. Goodchild in 1915. The discrepancy cannot possibly be attributed to its contamination by material from the sedimentary floor, since such a process would involve the addition of silica, alumina and alkalis, that is, precisely the reverse of what is observed to be the case. The anomaly is perhaps best explained by the presence of a small amount of accumulated olivine trapped in the sticky, rapidly congealing dolerite magma. The f-norm of the chill phase is 130, which according to T. Barth (1951) implies that on crystallisation pyroxene or olivine separates in advance of plagioclase. The slight deficiency of alumina and alkalis, and probably of titania as well, displayed by the chill phase may perhaps be attributed to the upward expulsion of part of the more mobile residue of the initial dolerite magma when subjected to a maximum degree of undercooling against the relatively cool sedimentary floor of the intrusion.


si
fm
c
al
alk
mg x 100
c-(al-alk) / fm
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A.
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0'
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2128'

Fig. 5

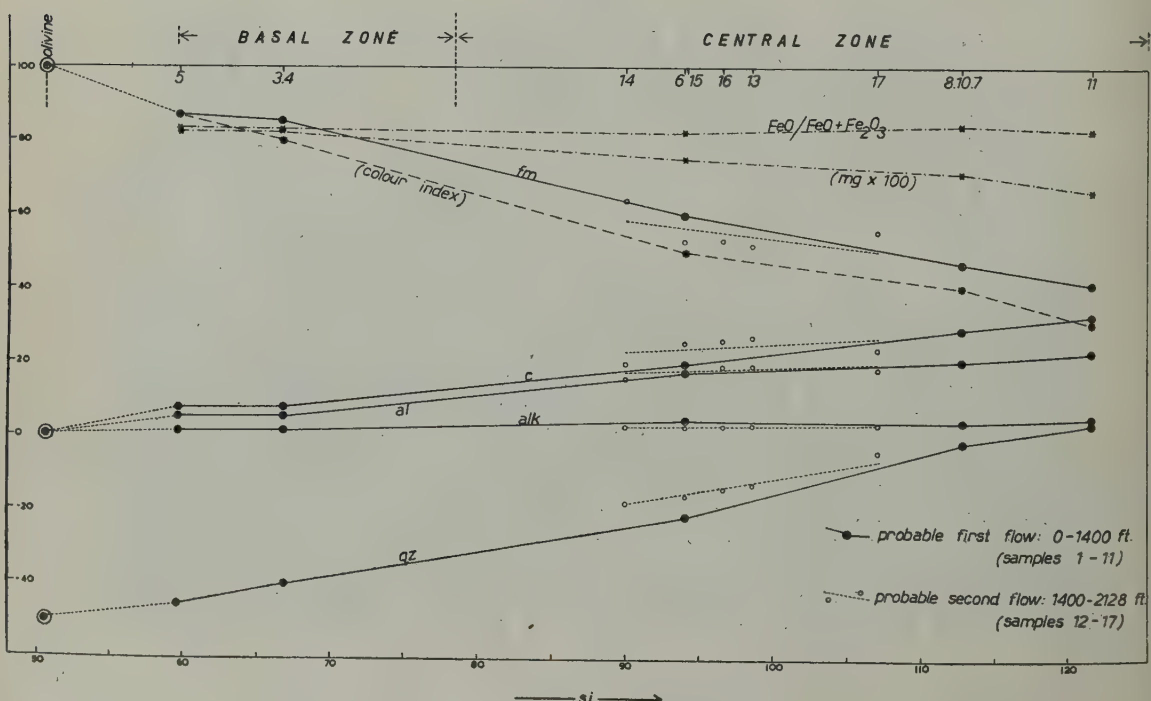
olivine
BASAL ZONE
CENTRAL ZONE
FeO/FeO+Fe2O3
(mg x 100)
(colour index)
fm
c
al
alk
qz
probable first flow: 0-1400 ft. (samples 1-11)
probable second flow: 1400-2128 ft. (samples 12-17)
si

Fig. 6

The Niggli values of the suite of analysed samples (Table IIb) when plotted against height (Fig. 5) or *si* as abscissa (see Fig. 6) demonstrate the fundamental antipathetic relation between the dominant *si* and *fm* molecular values throughout the succession. Mineralogically this feature is reflected by the abundance of olivine in the modes of the rocks with *si* less than 110 and peak or high *fm* values, while the degree of undersaturation is also disclosed by the negative *qz* curves in both diagrams. Attention should also be drawn to the almost constant and high *mg* ratios, and the fact that they rise sympathetically with *fm* indicates that the most magnesian olivine appears where the mineral is most abundant. That the olivine is of accumulative origin is also indicated by the trend of the *fm* and *qz* values towards, as well as the convergence of *al, c* and *alk* in, the direction of the relevant points indicated by the molecular values of olivine on the ordinate at *si* 50 (see Fig. 6). An increase in the percentage of modal olivine is also seen to be directly related to the melanocratic character of undersaturated rocks of increasing basicity because of the approximately parallel trend of the *fm* and the colour index curves in Fig. 6. In this graph too, the flat and constant $FeO/(Fe_2O_3 + FeO)$ ratio curve considered together with the increasing divergence of the *fm* and *mg* curves with a rise in *si* values, clearly indicates that ferrous iron enrichment is the dominant chemical differentiation trend displayed by the suite of rocks under consideration.

The fact that *alk* maintains a nearly constant value and *al* and *c* are almost sympathetic, while $al + c$ is markedly antithetic to *fm,* is portrayed in the mode primarily by the antipathetic relation of plagioclase feldspar and olivine (see Diagram I) when *qz* is strongly negative, and is indicative of the operation of in situ accumulative processes before the consolidation of the intrusion.

Though the most striking mineralogical variation is the antipathetic relation between olivine and plagioclase, it appears to be largely confined to the undersaturated sub-zones of the intrusion. In rocks with low negative or positive *qz* values, that is when the *si* value exceeds 110, a change in the mineralogical inter-relations is observed. With the disappearance of olivine its antithetic relation to plagioclase is first assumed by orthopyroxene but this association is temporary and is soon replaced by an analogous relation between plagioclase and clinopyroxene. Unlike olivine, orthopyroxene does not disappear, however, and though quantitatively less important than clinopyroxene it is still an important and constant mineral phase of the saturated and slightly oversaturated rocks of the succession under investigation. This would seem to imply that the excess of orthopyroxene in the immediate vicinity of the olivine-bearing sub-zones must be regarded as the product of an in situ reaction between a more siliceous rest-magma and olivine.

For the sake of convenience, the local irregularities portrayed by all the graphs between 720′ and 880′ or in the immediate vicinity of sample 9 may now be briefly discussed. The reappearance of olivine (Fo_{75} Fa_{25}) and the related local anomalous undulation of the modal curves appears to be a minor and probably impersistent feature of small lateral extent. It may be attributed to the accumulation of olivines of late arrival and origin, which were unable to penetrate the mush of plagioclase crystals which had already been differentially concentrated above the picrites. Theoretically such concentrates of olivines richer in iron, which may conceivably even give rise to

TABLE II

CHEMICAL ANALYSES

Sample	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17
	0′	12′	43′	185′	321′	361′	585′	716′	786′	1091′	1398′	1430′	1528′	1758′	2014′	2107′	2128′
SiO_2	52.16	45.31	44.50	42.50	41.01	47.61	52.14	51.03	45.65	51.01	52.40	50.32	48.64	47.59	47.08	47.65	50.59
Al_2O_3	13.55	8.39	7.00	5.29	5.83	14.82	14.61	15.26	7.99	17.13	16.44	17.34	15.46	14.14	16.24	15.90	14.25
Fe_2O_3	1.72	2.47	1.90	2.87	2.39	1.84	1.51	1.28	2.00	0.81	1.19	1.12	1.52	2.00	2.07	1.99	1.51
FeO	8.19	9.63	9.70	9.41	10.63	8.12	6.75	6.39	11.28	5.03	5.68	5.10	6.90	7.90	6.68	6.47	6.47
MgO	9.94	23.32	28.21	33.05	32.63	14.65	11.04	11.72	23.23	8.00	7.84	8.52	12.24	16.71	12.67	12.48	12.98
CaO	9.74	6.38	5.00	4.14	4.26	9.08	10.52	11.52	5.70	15.12	12.98	14.64	12.62	9.50	11.92	11.92	10.38
Na_2O	1.42	1.22	1.08	0.81	0.81	1.76	1.50	1.20	0.85	1.70	1.75	1.50	1.30	0.90	1.20	1.19	1.19
K_2O	0.58	0.43	0.38	0.33	0.35	0.50	0.55	0.38	0.55	0.42	0.40	0.30	0.28	0.27	0.31	0.32	0.40
H_2O-	0.15	0.13	0.13	0.12	0.15	0.15	0.05	0.02	0.05	0.03	0.05	0.04	0.02	0.04	0.10	0.07	0.07
H_2O+	1.68	1.35	1.46	1.45	1.70	1.23	1.20	0.92	1.49	0.83	1.10	0.75	0.88	0.98	0.83	0.87	1.35
TiO_2	0.90	0.76	0.60	0.52	0.35	0.65	0.62	0.52	0.70	0.48	0.50	0.32	0.32	0.40	0.55	1.20	0.55
P_2O_5	0.17	0.07	0.09	0.11	0.06	0.06	0.00	0.00	0.09	0.00	0.00	0.03	0.00	0.00	0.12	0.00	0.00
MnO	0.18	0.19	0.20	0.19	0.19	0.17	0.17	0.16	0.17	0.12	0.16	0.12	0.14	0.18	0.37	0.28	.042
S	—	0.66	—	—	—	—	—	—	0.34	—	—	—	—	—	—	—	—
Total	100.38	100.31	100.25	100.79	100.36	100.64	100.66	100.40	100.09	100.67	100.49	100.10	100.32	100.61	100.14	100.34	100.16
S-O	—	0.33	—	—	—	—	—	—	0.17	—	—	—	—	—	—	—	—
Total	—	99.98	—	—	—	—	—	—	99.92	—	—	—	—	—	—	—	—

NIGGLI VALUES

	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17
si	120.4	77.7	71.0	62.3	59.7	94.0	115.3	109.3	78.8	113.5	121.6	110.8	98.5	89.8	94.1	96.5	107.0
al	18.3	8.6	6.6	4.7	4.9	17.0	19.0	19.3	8.1	22.4	22.4	22.5	18.6	15.6	18.9	18.9	17.7
fm	53.6	77.4	82.9	87.6	86.9	59.6	51.9	51.4	79.3	37.5	40.8	39.5	51.3	63.1	52.8	52.5	55.7
c	24.1	11.6	8.5	6.4	6.6	19.3	24.9	26.3	10.6	36.0	32.3	34.5	27.3	19.2	25.5	25.9	23.7
alk	4.0	2.5	2.0	1.4	1.4	4.0	4.0	3.0	2.1	4.1	4.5	3.5	2.9	2.0	2.6	2.8	2.9
k	0.21	0.21	0.19	0.19	0.19	0.15	0.20	0.17	0.30	0.13	0.13	0.11	0.13	0.17	0.14	0.14	0.17
mg	0.64	0.78	0.81	0.83	0.82	0.73	0.71	0.73	0.76	0.71	0.67	0.71	0.73	0.75	0.72	0.72	0.74
c/fm	0.45	0.15	0.10	0.07	0.08	0.32	0.48	0.51	0.13	0.96	0.79	0.87	0.53	0.30	0.48	0.49	0.43
qz	+4.4	−32.3	−37.0	−43.3	−45.9	−22.0	−0.5	−2.7	−29.6	−2.9	+3.6	−3.2	−13.1	−18.2	−16.7	−14.7	−4.6
h	12.9	7.7	7.7	7.1	8.2	8.5	9.1	6.4	8.6	5.9	8.5	5.4	6.1	6.3	5.3	5.9	9.5
ti	1.5	1.0	0.8	0.5	0.4	0.9	1.1	0.8	0.9	0.9	0.8	0.5	0.5	0.6	0.8	1.8	0.9
p	0.1	0.1	0.1	0.1	0.1	0.1	0.0	0.0	0.1	0.0	0.0	0.0	0.0	0.0	0.1	0.0	0.0

NORMS

	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17
Q	4.68	—	—	—	—	—	1.92	0.60	—	—	3.42	—	—	—	—	—	—
or	3.34	2.78	2.22	1.67	1.67	2.78	3.34	2.22	3.34	2.22	2.22	1.67	1.67	1.67	1.67	1.67	2.22
ab	12.05	9.96	8.91	6.81	4.72	14.67	12.58	9.96	7.34	14.15	14.67	12.58	11.00	7.86	9.96	9.96	9.96
an	28.63	16.12	13.34	10.01	11.40	31.14	31.41	35.31	16.12	38.09	35.86	39.75	35.58	33.36	37.81	37.25	32.25
ne	—	—	—	—	1.14	—	—	—	—	—	—	—	—	—	—	—	—
di $CaSiO_3$	7.89	6.15	4.41	3.94	3.71	5.57	8.70	9.05	4.76	15.43	11.95	13.69	11.25	5.80	8.58	9.16	8.12
di $MgSiO_3$	4.90	4.50	3.30	3.00	2.80	3.10	5.80	6.20	3.40	10.20	7.70	9.20	7.60	4.10	5.90	6.50	6.30
di $FeSiO_3$	2.51	1.06	0.66	0.53	0.53	2.24	2.24	2.11	0.92	4.09	3.43	3.43	2.77	1.19	1.98	1.85	0.92
hy $MgSiO_3$	19.90	12.30	8.80	3.70	—	11.40	21.80	23.10	17.40	9.30	11.90	12.10	10.90	18.50	11.70	14.30	25.90
hy $FeSiO_3$	10.03	2.90	1.98	0.66	—	3.17	8.32	8.05	5.28	3.83	5.41	4.62	4.09	5.54	3.70	3.96	9.64
ol Mg_2SiO_4	—	28.98	40.88	53.13	55.16	15.47	—	—	26.11	0.35	—	—	8.47	13.44	9.87	7.28	0.14
ol Fe_2SiO_4	—	7.44	10.00	10.30	12.95	5.71	—	—	8.67	0.10	—	—	3.36	4.49	3.57	2.24	0.01
Fe_7S_8	—	0.88	—	—	—	—	—	—	0.44	—	—	—	—	—	—	—	—
mt	2.55	3.48	2.78	4.18	3.48	2.55	2.09	1.86	3.02	1.16	1.86	1.62	2.09	3.02	3.02	3.02	2.09
il	1.67	1.52	1.06	0.91	0.61	1.22	1.22	0.91	1.37	0.91	0.91	0.61	0.61	0.76	1.06	2.28	1.06
ap	0.34	0.34	0.34	0.34	0.34	0.34	—	—	0.34	—	—	—	—	—	0.34	—	—
H_2O	1.83	1.48	1.59	1.57	1.85	1.38	1.25	0.94	1.54	0.86	1.15	0.79	0.90	1.02	0.93	0.94	1.42
Total	100.32	99.89	100.27	100.75	100.36	100.74	100.67	100.31	100.05	100.69	100.48	100.06	100.29	100.75	100.09	100.41	100.03

MODES

	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17
Quartz	n.d.	—	—	—	—	—	1.40	0.4	—	—	0.55	—	—	—	—	—	—
Alk. Feldspar	n.d.	0.1	0.15	0.05	0.15	0.43	1.82	0.4	0.6	0.5	0.40	0.05	0.16	0.05	0.06	0.12	0.11
Plagioclase	n.d.	30.0	22.0	18.0	13.0	49.0	57.5	59.0	28.0	59.0	60.5	64.0	52.0	49.0	60.0	53.5	52.0
Clinopyroxene	n.d.	7.0	8.0	7.5	10.5	10.5	12.0	21.0	9.0	34.0	27.0	23.5	23.0	12.5	15.0	20.0	17.0
Orthopyroxene	n.d.	24.5	16.0	9.5	11.0	13.0	25.0	18.0	25.0	7.0	11.0	7.0	8.0	11.0	9.0	9.0	27.0
Olivine	n.d.	34.0	50.5	62.5	63.0	24.5	—	—	33.0	tr	—	5.5	16.0	26.5	15.0	17.0	3.0
Apatite	n.d.	n.d.	n.d.	n.d.	0.1	0.2	0.3	0.06	0.1	0.005	0.02	n.d.	0.1	0.04	0.08	0.05	n.d.
Ore	n.d.	0.9	1.3	1.2	1.1	0.6	0.5	0.4	1.0	0.2	0.3	0.0	0.2	0.45	0.4	0.4	0.4
Biotite	n.d.	3.5	2.5	1.1	1.7	2.1	1.1	0.6	3.0	0.4	0.3	0.0	0.2	0.5	0.4	0.2	0.05
Hornblende	n.d.	—	—	—	—	—	1.8	—	—	—	—	—	—	—	—	—	—
Colour Index	n.d.	70	78	82	87	50	39	40	72	41	39	36	48	51	40	47	48

1. Chill phase 2. Olivine hyperite 3, 4, 5. Picrite 6. Olivine hyperite 7. Quartz hyperite 8. Quartz hypersthene gabbro 9. Olivine hyperite 10. Hypersthene gabbro 11. Quartz hypersthene gabbro 12, 13, 14, 15, 16. Olivine hypersthene gabbro 17. Olivine hyperite

local banding on a small scale and at different elevations, are not believed to be of any real quantitative significance, as far as the main differentiation trend is concerned.

The zoning of the plagioclase in the picrites, the slightly lower anorthite content and the poikilitic habit are attributed to small amounts of upward migrating rest-magma residual from the in situ crystallisation of olivine trapped by the main gravitative shower of early magnesian olivine.

If we disregard the values of the clearly abnormal sample 9 in Figure 5, it will be seen that *c* is slightly higher but generally speaking sympathetic to *al* and antithetic to *fm*. Where *fm* is lowest the difference between *c* and *al* attains a maximum, and *c-(al-alk)* is therefore still greater than *al*. The sympathetic trend of *c-(al-alk)/fm* relative to the augite/olivine plus pyroxene ratio indicates that this excess of lime not utilised in the formation of basic plagioclase is incorporated in the substance of the clinopyroxene in the olivine-free sub-zones of the intrusion. This relation is also borne out by the fact that *al-alk* rises less steeply than *c* with increasing *si* values in Figure 5.

Above the 1,400-foot level olivine again becomes an important constituent of the hypersthene gabbro of the Central Zone and in these rocks the mutual relations of the constituent minerals display many features in common with the olivine-richer rocks lower in the succession. The lower concentration of olivine, its irregular vertical distribution and slightly higher fayalite content, as well as the relatively larger proportion of clinopyroxene, are among the most important differences displayed by the upper olivine-bearing subzone (See Diagram I). It will be observed that at 2,128 feet, that is on the highest attainable level in the section of the intrusion under consideration, there is a sharp reduction in the olivine content of the rock and its disappearance is again accompanied by a disproportional increase in the amount of orthopyroxene. Apart from the anomaly already referred to, the rocks of the Central Zone between the upper and lower olivine horizons are characterised by the presence of small amounts of alkali feldspar, quartz, and slightly less calcic plagioclase than is present in the overlying olivine hypersthene gabbros.

From the foregoing it would appear that the 1,400-foot level represents the top of the earliest and main inflow of magma, whereas only the lower portion of the later and probably less completely differentiated and consolidated magmatic heave is now visible in the Waterfall Gorge section of the intrusion. Geomorphologically the top of the 1,400-foot thick flow is probably represented by the undulating crest of the precipitous escarpment flanking the Insizwa Range.

It is believed that the second influx of magma occurred before the complete consolidation of the more acid differentiate of the first inflow. This would account for the lack of invasive evidence, the absence of concentrations of olivine in the central quartz-bearing hypersthene gabbro, and the presence of segregations of olivine-free gabbro pegmatite just above the precipitous cliffs. The apparent absence of acid differentiates which characterise the roof zone at Ingeli or the anastomosing vein-like segregations of micropegmatite described by other investigators in the vicinity of the Insizwa and Nolangeni peaks also seem to indicate that the magma of the

second flow mixed with and incorporated much of the substances of the acid differentiate of the first flow.

When one considers the position of the studied section with reference to the form of the original intrusion as postulated by D. L. Scholtz (1936), it would seem that the average composition of the lower 1,400 feet of the intrusion should prove to be more basic than the average composition of the Karroo dolerite magma.

This conclusion is based on the assumption that trough sections, in contra-distinction to ridge sections, of the undulatory sheet-like intrusion would be enriched in olivine and depleted in acid differentiates and vice versa. Since the Waterfall Gorge section intersects the trough-rim characterised by the presence of a rapidly thickening concavo-convex picrite body, it is to be expected that the average calculated composition of the almost intact lower flow should approximate that of the neighbouring dolerites, plus some olivine.

Table III.

	I	II	III		IV	V	VI		VII	VIII
SiO_2	50.9	52.5	49.1	Q	2.3	—	—	olivine	15.2	6.1
Al_2O_3	15.7	15.4	13.8	ol	—	3.4	12.9	plagioclase	47.9	47.5
FeO	9.4	9.3	7.0	or	5.0	3.3	2.2	orthopyr.	13.3	38.8
Fe_2O_3	0.9	1.2	1.5	ab	17.8	16.8	9.4	clinopyr.	19.3	
MgO	8.6	7.1	15.2	an	30.0	33.1	31.4	ore + biot.	1.8	3.6
CaO	10.8	10.3	10.7	di	16.7	16.1	17.5	qz + cryptop.	2.0	4.0
Na_2O	2.0	2.1	1.1	hy	24.3	24.4	23.0			
K_2O	0.6	0.8	0.4	mt	1.9	1.4	1.1			
TiO_2	0.9	1.0	0.5	il	2.0	1.7	1.1			
P_2O_5	0.1	0.1	0.0	ap	0.3	0.3	0.0			

I Average Kokstad type dolerite (13 analyses). F. Walker.
II Average Karroo dolerite. F. Walker.
III Weighted average composition of lower flow, Waterfall Gorge.
IV Norm of average Karroo dolerite.
V Norm average Kokstad type dolerite.
VI Norm of lower flow, Waterfall Gorge, Insizwa.
VII Weighted mode of lower flow, Waterfall Gorge, Insizwa.
VIII Mode of average Kokstad type dolerite. F. Walker.

Table III indicates that the planimetric weighted average chemical composition of the 1,400-foot thick lower flow at Waterfall Gorge differs markedly from that of the average Karroo Dolerite in silica and magnesia content, and to a lesser extent in ferrous iron, alumina and soda.

With the possible exception of silica, F. Walker's average dolerite of the Kokstad type shows variations of almost precisely the same nature and magnitude. When the norms of these three average analyses (calculated water-free) are compared, the differences mentioned above are reflected in the relative proportions of albite and olivine, thereby corroborating the conclusions previously arrived at about the addition of olivine and the probable loss of alkali feldspar and quartz. Allowing for the variability in the chemical composition of the mineral phases, the weighted planimetric mode of the flow based on the integration of 35 thin-sections, agrees fairly well with the norm and differs in an essentially analogous manner from F. Walker's average mode

of the Kokstad type dolerite. In this respect it is unfortunate that no mode of the average Karroo dolerite is available for comparison.

The second magma flow is thought to have been emplaced some 1,400-feet above the base of the intrusion as a result of the subsidence of the floor, as has been briefly mentioned. It will be recalled that there is reason to believe that only the lower 750 feet of the rock succession, representing the second magmatic heave, have been preserved from denudation in the Waterfall Gorge section of the Insizwa intrusion. The modes of the rocks from different elevations (see Diagram I) suggest that the differentiation of the second flow was less complete than that of the earlier flow. Residual ferrous iron from the consolidating lower flow may have been responsible for the olivines being richer in fayalite as well as for their range of crystallisation being overlapped by that of plagioclase. This would account for the oscillatory but complementary olivine-plagioclase association, or rude banding of the rocks, owing to the mutual interference of showers of those minerals accumulating along a gravity gradient, as suggested by the experiments of R. R. Coates (1936) and by the general trend of the differentiation shown by the Niggli values of the rocks in Figures 5 and 6.

The Waterfall Gorge section of the Insizwa intrusion does not include any acid or basic rocks of extreme composition such as are known to exist in that region, and does not therefore provide a satisfactory basis for investigating the complete differentiation trend. For purposes of comparison, however, the new analyses have been utilised in the preparation of Niggli's Q.L.M. diagram (Fig. 8) and the illuminating but simple ternary diagram with alkalis, ferrous iron and magnesia as variables (Fig. 7).

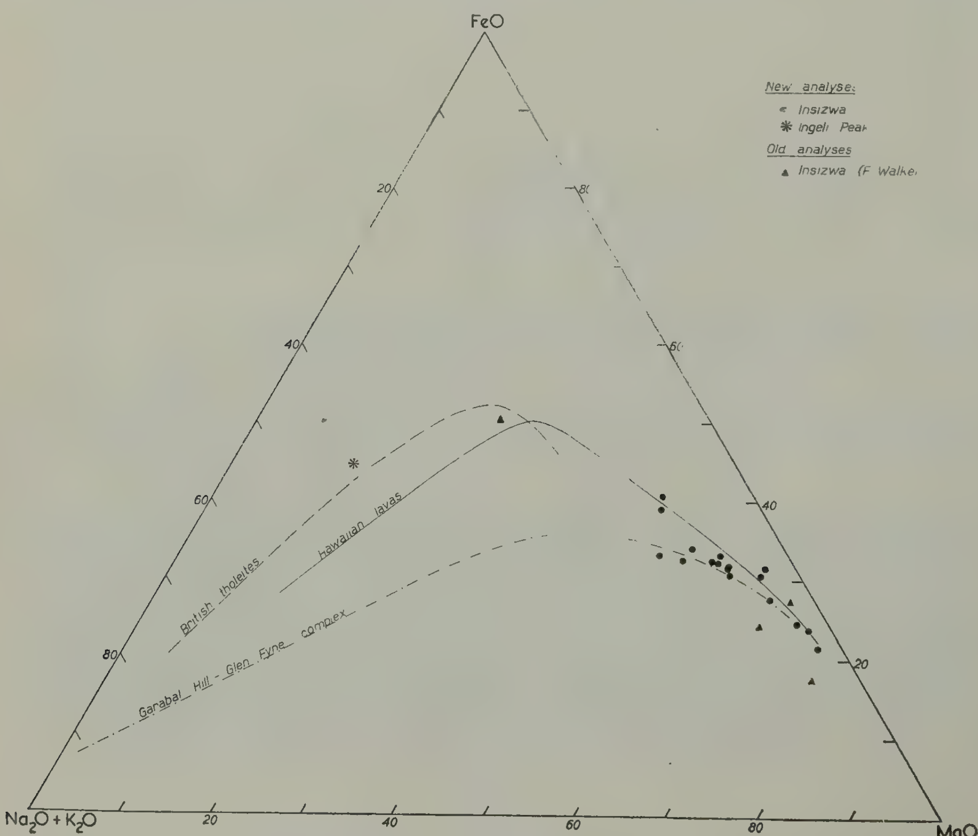

Fig. 7

The early differentiation trend, as previously indicated, displays a distinct tendency towards ferrous iron enrichment and simultaneously slight increase in alkali content, thereby corroborating the mineralogical evidence of the much stronger fractionation of the mafic than of the felsic phases in the section of the intrusion under consideration. If a superior old analysis of a rock more basic than any of the present suite is included (solid triangle), then the more magnesian character of some extreme early differentiates is accentuated. This feature can also readily be accounted for, by the more pronounced crystal fractionation of the ferro-magnesian constituents during the initial stages of the differentiation process.

If the analyses of Insizwa differentiates which are more acid (solid triangles) than those appearing in the Waterfall Gorge section, are considered, it will be seen that the differentiation trend is in the direction of the crest of the Hawaiian curve, beyond which region there is reason to believe that the correspondence in trend will be less evident.

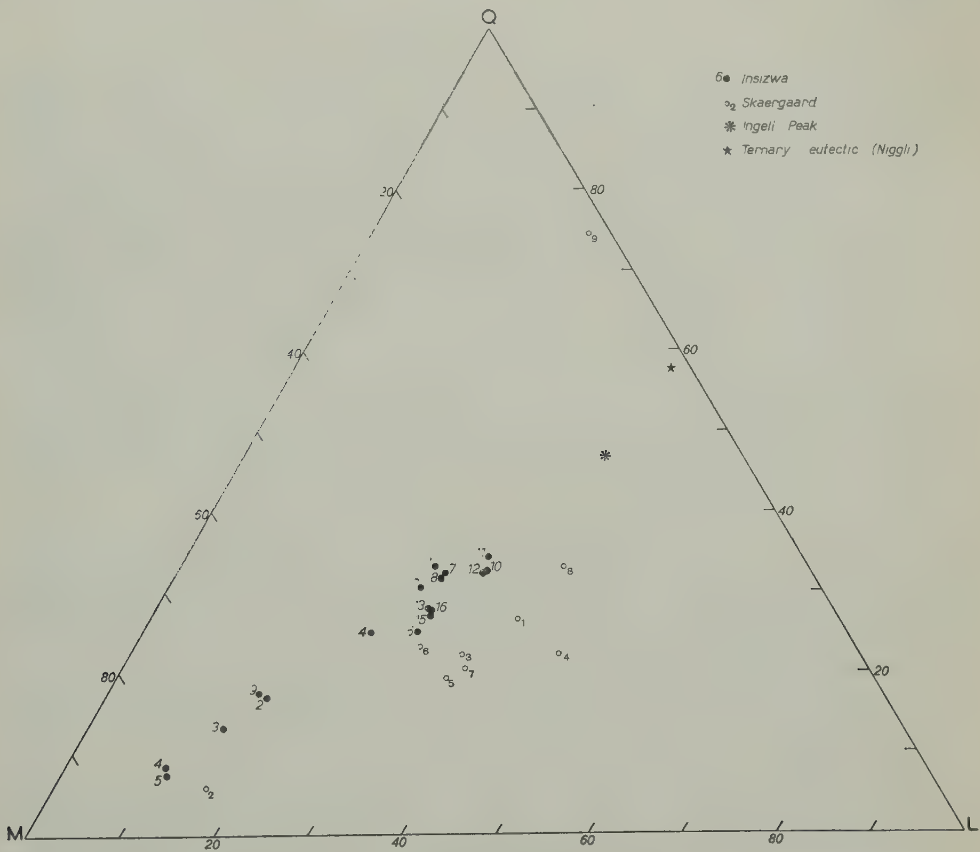

Fig. 8. Q-L-M values of the Insizwa samples compared with the 9 samples chosen by L. R. Wager and R. L. Mitchell (1943) as being representative of the order of differentiation of the Skaergaard.

In short, as far as the magnesia: ferrous iron: alkali ratio is concerned, the Insizwa rocks on the one hand resemble those of the Hawaiian petrographic province in differentiation trend during the early stages, and on the other hand those of the British Tholeiitic province during the later stages of the differentiation process.

Figure 8 indicates that in the case of the basic rocks of the thoroughly investigated Skaergaard intrusion and the Insizwa, accumulative processes involving the separation of mafic and felsic minerals was a significant feature of the differentiation. In this Q.L.M. diagram the more feldspathic character of the Skaergaard type rock field relative to the Insizwa field, and the parallel trend of the differentiation processes are well illustrated. With increasing acidity, an abrupt change away from the direction of Niggli's ternary eutectic (★) distinguishes the ultimate highly quartzose granophyric Skaergaard differentiate ($^{\circ}_{9}$) from the far less acid product characterising the roof zone at Ingeli (*).

V. PETROGENESIS

The correlation of the gabbroidal sheet comprising the Insizwa, Tonti, Tabankulu and Ingeli masses with the multitudinous Karroo dolerite intrusions has been firmly established by the investigations of A. L. du Toit and D. L. Scholtz. As a result of their publications, it has been generally accepted that the petrographic variations encountered across these masses originated from a process of fractional crystallisation modified by gravitational differentiation. D. L. Scholtz produces evidence to indicate that the injection of basaltic magma was accomplished in at least two successive but almost synchronous heaves and that crystallisation proceeded in an effectively closed system under a pressure of at least 750 atmospheres. The salient features of the differentiation comprise the settling and accumulation of olivine crystals in the lowest portions of the undulating sheet, together with the upward migration and concentration of a relatively acid, volatile-rich, late-liquid fraction towards the arched crests of the intrusive body, this process giving rise to the broad division of the intrusion into three distinct petrological units: the Roof Zone, the Central Zone and the Basal Zone.

The present investigation of the Waterfall Gorge section of the Insizwa shows that the detailed petrographic variations are adequately accounted for by the igneous history which has been proposed for the intrusive as a whole. However, certain features do at first appear to be in variance with the ideal sequence to be anticipated from such a process.

1. The amount of olivine in the section (15% by volume) is considerably larger than would be expected from the tholeiitic composition of the average Karroo dolerite magma, even assuming a very complete differentiation.

2. The complementary acid differentiate is entirely lacking from the section.

3. The alternation of three olivine-bearing zones with two olivine-free zones indicates some departure from the ideal and undisturbed gravitative crystallisation such as is displayed by the single olivine-rich layer in the Palisade diabase sill.

4. The greatest concentration of olivine, within the picrite of the Basal Zone, is found at a height of 300 to 350 feet, above which there is a fairly abrupt decrease and disappearance of this phase. From the hypothesis advocating the settling of early-formed olivine through a fluid magma on to a solid contact floor, it would be expected that the percentage of olivine would be greatest at the base and would gradually diminish upwards.
5. Assuming that the lower horizons contain crystals of earlier formation than the higher, the ferromagnesian minerals should be most magnesian at the base and become progressively ferrous upwards. In the picrite, however, the most magnesia-rich minerals are found in the centre of this sub-zone and tend to become more iron-rich both downwards as well as upwards.
6. Fairly abrupt changes in mineral composition and relative proportions are encountered in the central olivine hyperite and at the base of the upper hypersthene gabbro.

These features may, however, all be resolved by the introduction of some modifications in the normal process of magmatic differentiation and by a consideration of the peculiarities of the particular section. The Waterfall Gorge profile is located within the rim of one of the depressions in the sheet and, as may be expected from the configuration of the sheet, it is comprised of a well-developed Basal Zone and probably a complete section of the Central Zone, but lacks any material from the upper acid phase or Roof Zone. The absence of the troctolite layer, which demarcates the top of the Basal Zone in the Ingeli and along the Umzimhlava section of the Insizwa, also supports the conclusion that the profile is not across the deepest portion of the depression, where the development of the troctolite phase is expected to be greatest. It should perhaps also be emphasised that both the importance and the extent of lateral variations cannot be estimated from material obtained within the confines of a single profile. In this respect, attention has already been directed to the anomalous nature of the relatively thin central olivine-bearing sub-zone and reasons have been given for supposing that this layer is probably of insignificant lateral extent. Thus, the prominence of this sub-zone in the Waterfall Gorge section may be largely fortuitous.

In the preceding chapter it has been shown that there is reason to believe that the rock types of this profile crystallised from two successive influxes of magma, the second being injected after sufficient time had elapsed for the crystal settling of the first to have reached a fairly complete stage. The first emplacement, which exceeded a thickness of 1,400 feet in this section, represents the main influx of magma into the intrusion. Along the immediate contact with the relatively cool country rock, this magma experienced an abrupt chilling and resultant solidification to form the fine-grained contact phase. A consideration of the chemical composition of this chilled phase indicates that the initial magma, at the beginning of its emplacement, possessed a tholeiitic composition close to that of the average Karroo dolerite. It must be recognised, however, that the composition of the magma could change during emplacement, and the average composition and mineralogy of the rest of the section points to the bulk of the intrusion being of olivine basalt. Apparently this is not an uncommon feature for Walker and Poldervaart (1949, p. 622) note that tholeiite occurs along the borders of many of the

olivine dolerite intrusions of the Karroo, and they indicate that the composition of marginal olivine dolerite magma was modified to a tholeiite by the solution of hyperfusible components from the country rock. However, no significant assimilation of wall-rock material by the magma after intrusion is to be expected, for such a reaction would be endothermic and would tend to hasten the chilling process. It is noteworthy that the Northern Transkei is characterised by abundant intrusions of dolerites of the "Kokstad type", which is notably more olivine-rich than the average Karroo Dolerite. It is also significant that this region is situated near the deepest portion of the Karroo Basin and according to Walker and Poldervaart (p. 684) occupies the focal area of the igneous activity. It is therefore not unreasonable to expect that the bulk of this initial intrusion was of primary olivine basalt and that the first influx, contaminated in depth by the incorporation of sialic material, possessed a more tholeiitic character.

After the chilling of the basal contact phase, continued loss of heat to the underlying country rock resulted in the crystallising of a layer of magma immediately above the chill phase. At first this solid layer extended rapidly upwards but its rate of growth gradually decreased as this zone became thicker and as the country rock acquired a higher temperature. In the meantime, olivine crystals were separating from the magma throughout the rest of its extent and were settling towards the base of the intrusion, where they were incorporated within the congealing border facies to form the basal olivine hyperite.

As the rate of the upward extention of this solid floor decreased, a greater proportion of the sunken olivine crystals were incorporated within it, thereby accounting for the progressive upward enrichment in olivine across the basal olivine hyperite and picrite. The lower portion of the intrusion, from which the earliest olivine crystals separated and settled, probably experienced a rapid initial cooling during emplacement, as a result of which the magma temperature may have been depressed below the liquidus of the thermal curves of the olivine isomorphous series. If this assumption is correct, the first olivines would therefore be more fayalite-rich than those which would normally separate from the same mixture under more gradual conditions of cooling. With the progressive adjustment to equilibrium, owing partly to the rise of magma temperature from latent heat released by crystallisation, more forsterite-rich olivines would separate, these conditions prevailing throughout the major crystallisation period of this mineral. In the upper half of the picrite, however the normal trend of fractional crystallisation is restored with the settling of progressively more iron-rich olivines.

In contrast to the olivine, the other essential minerals of the picrite, namely plagioclase and pyroxene, appear to have crystallised in situ in the form of large poikilitic plates, while an acid fraction of the liquid, charged with volatiles, was displaced upwards to be concentrated in the lower quartz hyperite.

After the base of the intrusion had solidified to a height of about 350 feet, a point was rather suddenly reached at which the remaining liquid had become so changed as to be unable to precipitate any more olivine. A large amount of the olivine accumulated either on the floor or in the process of settling, was partially or completely resorbed by the magma in the peritectic stage of reaction to form pyroxene. This change in the

crystallising phase resulted in the replacement of olivine by orthopyroxene as the principal mafic mineral throughout the lower olivine hyperite as well as the hypersthene gabbro. In this respect, the close correspondence in both petrology and mineralogy between the central and the lower olivine hyperites is noteworthy. A possible explanation of this somewhat anomalous layer may therefore be that it represents the material from an originally dense shower of olivine from magma occupying the adjoining but now eroded, dome in the sheet.

The textural relationships between the essential minerals of the lower quartz-hyperite and the hypersthene gabbro are instructive. It is noted that the textures of the pyroxenes range from ophitic at the base, through sub-ophitic, to prismatic at the summit. This indicates that in the lower levels pyroxene succeeded the bulk of the plagioclase in the crystallisation sequence, possibly deriving much of its material from the resorption of olivine. At higher levels, however, the two minerals seem to have crystallised simultaneously and independently.

In this sub-zone too, the gradual increase in the ratio of augite to orthopyroxene is noted and is reflected by a rising *c/fm* ratio, undoubtedly owing to the early separation and removal of ferromagnesian components. As a result of this enrichment in lime relative to magnesia and ferrous iron, orthopyroxene did not separate as a primary precipitate in the overlying central quartz-bearing hypersthene gabbro. Instead, plagioclase was accompanied mainly by augite as the first crystallising phases, forming a mush of fairly uniform packing. The gradually increasing albite content of the plagioclase as well as the slight rise in the FeO/MgO ratio of the augite from the base to the top of this sub-zone, shows that some downward settling of these products, with a complementary upward migration of rest liquid, took place. Interstitial magma, remaining in the mush, crystallised later as the pseudo-poikilitic orthopyroxene, followed by alkali feldspar and quartz.

At about this stage a fresh influx of magma was emplaced between the almost completely consolidated central quartz-bearing hypersthene gabbro and the roof of the intrusion, as a result of the further settling of the floor. In the section under consideration, it was of smaller bulk than was the first, but, while following the undulatory habit of the original sheet, it was probably of highly variable thickness. The initial composition of this second magmatic influx was probably very similar to that of the first, but it was modified by mixing with the still liquid fraction of the original intrusion, which undoubtedly showed enrichment in ferrous iron, alkalis and silica. The new cycle of fractional crystallisation was initiated with the precipitation of plagioclase, followed shortly by olivine and later by augite. The more or less rhythmic variation and complementary nature of these minerals in the upper olivine hypersthene gabbro suggest that a differential settling process was operative, as has been postulated by R. R. Coates (1936), to account for the rhythmic layering in mafic rocks. Orthopyroxene is again of relatively late formation and the pseudo-poikilitic aggregates of this mineral were probably precipitated by reaction between the interstitial magma and the olivine of the crystal mush.

The accumulation of these early crystal phases in this section again indicates that the profile is located in or near a depression in the sheet. It is unfortunate that the absence from this profile of the late acid differentiate, which presumably migrated to the higher domes of the intrusion, has limited

this investigation to the earlier stages of the differentiation process. However, the present study is the first of a series of detailed investigations along specific profiles across the intrusion. At least two further profiles across the Insizwa mass and three crossing the very complete section preserved in the Ingeli range are being investigated by senior students and members of this department. Further investigations are also being conducted on the economic geology of the associated magmatic ore deposits and it is hoped that a more complete picture of the igneous history of this fascinating intrusion will evolve.

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PLATE 1

Pm. 1 Basal contact with hornfels. Note radiating texture of pyroxene aggregates in the chill-phase. Ordinary light, × 12 (borehole W.G. 5).

Pm. 2 Vertical section showing preferred sub-horizontal (photograph rotated through 90°) arrangement of plagioclase and pyroxenes, near the top of the lower quartz hypersthene gabbro. Crossed nicols, × 3.5 (borehole W.G. 5, 716-foot level).

Pm. 3 Pigeonite in twinned augite in the basal hyperite. Crossed nicols, × 60 (borehole W.G. 5, 1-foot level).

Pm. 4 Pseudo-poikilitic texture of orthopyroxene in central and upper sub-zones. Note twinned augite inclusions in optical continuity in orthopyroxene in lower half of photomicrograph. Crossed nicols, × 30 (1,855 feet above contact).

Pm. 5 Twinned olivine in picrite. Crossed nicols, × 60 (borehole W.G. 5, 85-foot level).

Pm. 6 Complex zoned plagioclase, typical of the Insizwa rocks. Note resorption of the oscillatory zoned core and the simultaneous extinction of the innermost core and the mantle. Crossed nicols, × 60 (borehole W.G. 5, 351-foot level).

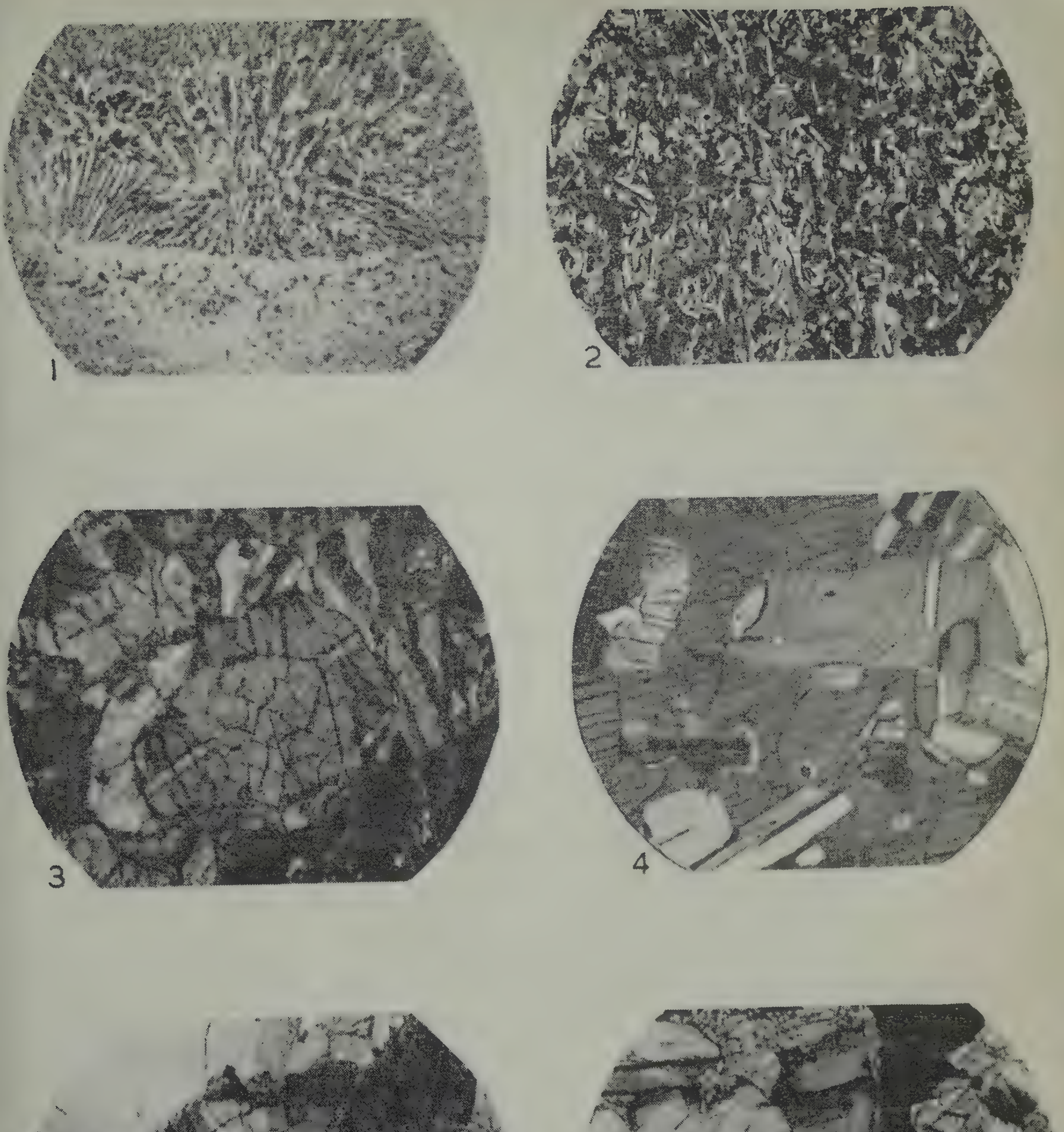


PLATE I.

PLATE II

Pm. 7 Pseudo-twinning lamellation displaying patchy extinction in a poikilitic orthopyroxene. Crossed nicols, × 105 (borehole W.G. 5, 127-foot level).

Pm. 8 Same section, rotated through 6° to obtain extinction of patches previously light. The included augite and olivine remain bright. Crossed nicols, × 105 (borehole W.G. 5, 127-foot level).

Pm. 9 Pseudo-twinning lamellation displaying patchy extinction in a poikilitic orthopyroxene. The extinction angle between adjacent lamellae is larger than in the case of pm. 7 and 8. Crossed nicols, × 18 (borehole W.G. 5, 36-foot level).

Pm. 10 Same section, rotated through 12° to obtain extinction of patches previously light. Crossed nicols, × 18 (borehole W.G. 5, 36-foot level).

Pm. 11 Well developed clinopyroxene intergrowth parallel to the pseudo-twinning lamellation of a pseudo-poikilitic orthopyroxene. Compare Fig. 4(b). Crossed nicols, × 17 (1,757 feet above contact).

Pm. 12 Herring-bone pattern of clinopyroxene platelets in a pseudo-poikilitic hypersthene. Compare Fig. 4(e). Crossed nicols, × 60 (1,091 feet above contact).

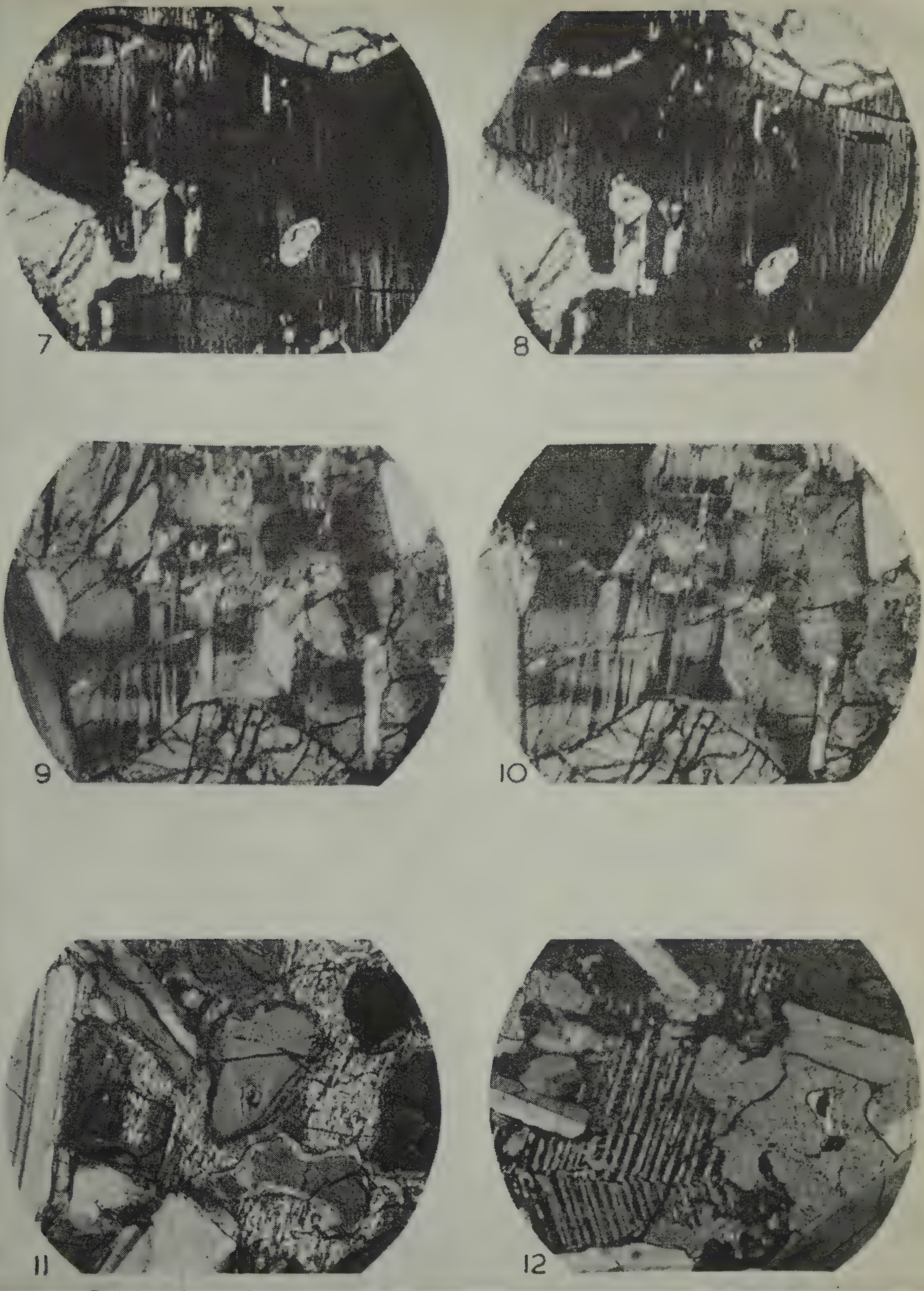
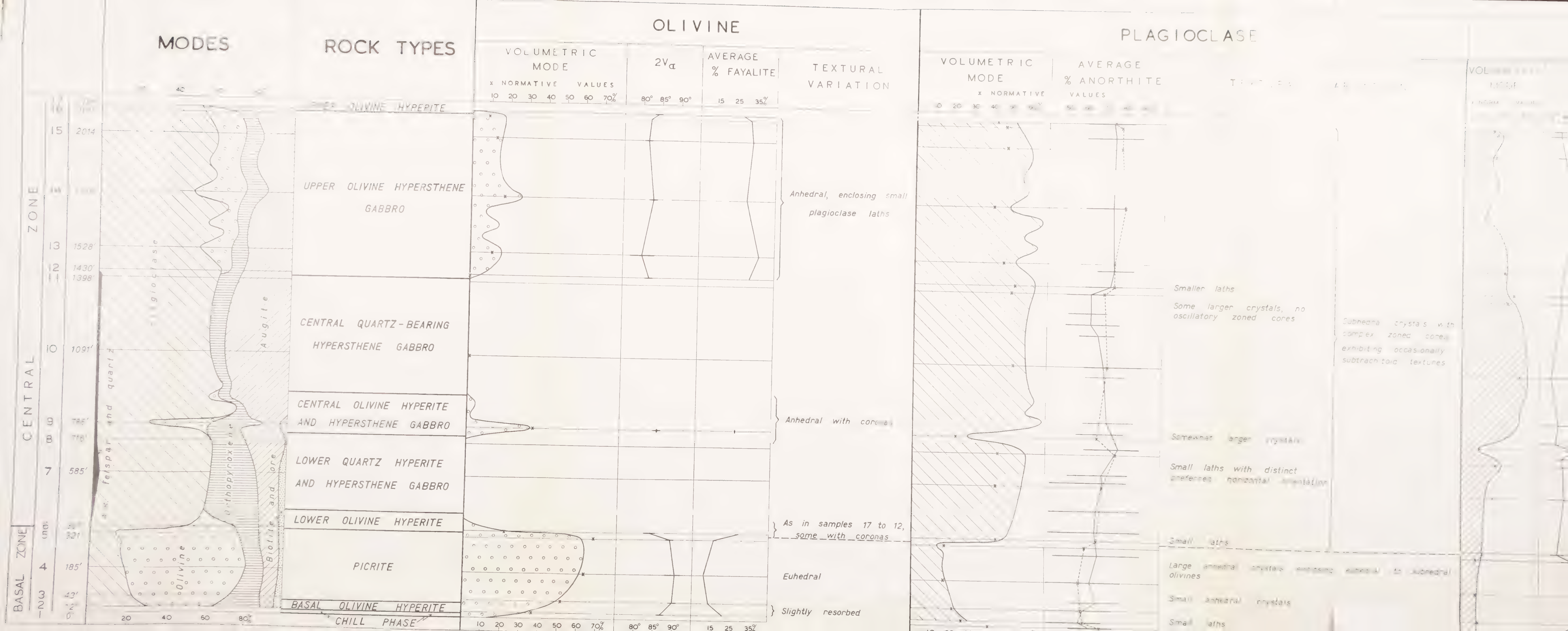
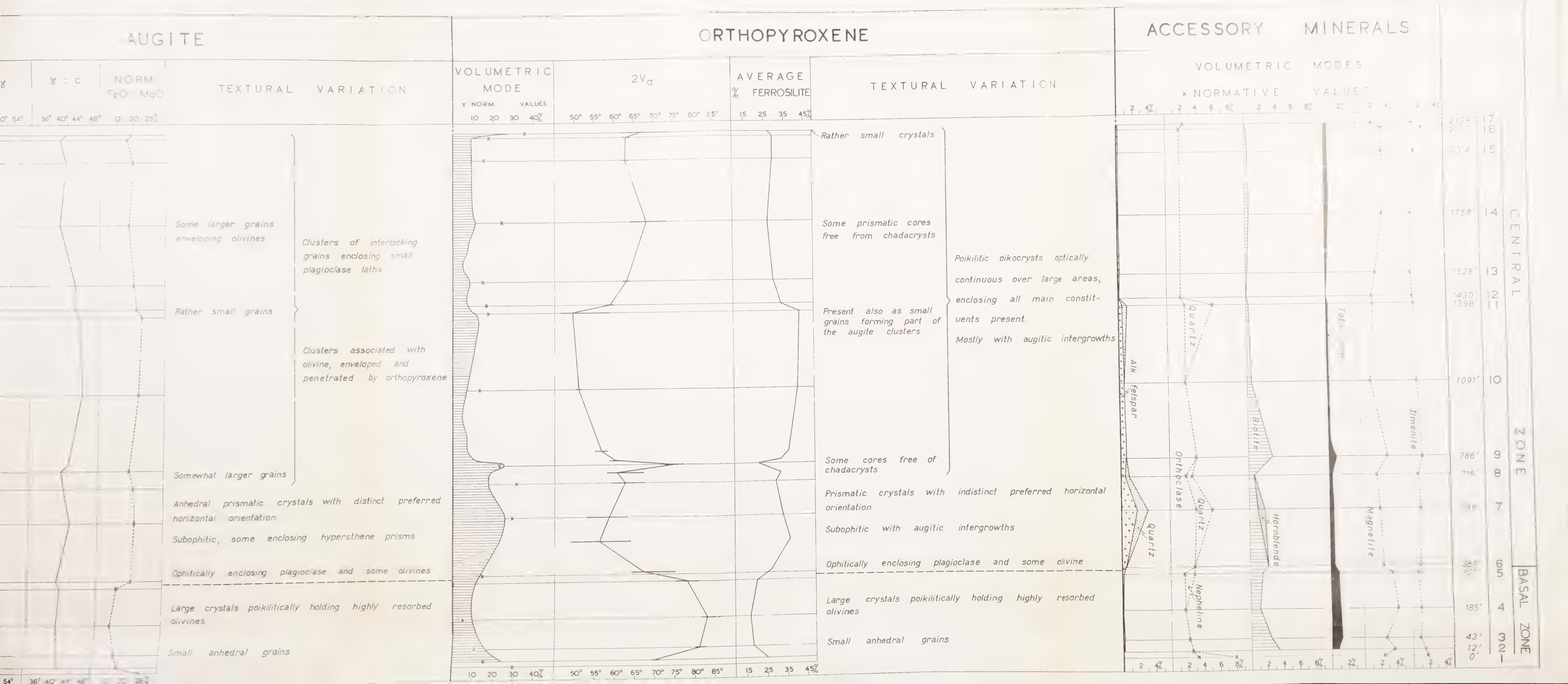


PLATE II.


MODES
ROCK TYPES
OLIVINE
VOLUMETRIC MODE
x NORMATIVE VALUES
10 20 30 40 50 60 70%
2Vα
80° 85° 90°
AVERAGE % FAYALITE
15 25 35%
TEXTURAL VARIATION
PLAGIOCLASE
VOLUMETRIC MODE
x NORMATIVE VALUES
AVERAGE % ANORTHITE
VALUES
CENTRAL ZONE
BASAL ZONE
15 2014'
13 1528'
12 1430'
11 1398'
10 1091'
9 785'
8 715'
7 585'
4 185'
3 43'
0'
20 40 60 80%
plagioclase
augite
orthopyroxene
olivine
Biotite and ore
alk. felspar and quartz
UPPER OLIVINE HYPERSTHENE GABBRO
CENTRAL QUARTZ-BEARING HYPERSTHENE GABBRO
CENTRAL OLIVINE HYPERITE AND HYPERSTHENE GABBRO
LOWER QUARTZ HYPERITE AND HYPERSTHENE GABBRO
LOWER OLIVINE HYPERITE
PICRITE
BASAL OLIVINE HYPERITE
CHILL PHASE
Anhedral, enclosing small plagioclase laths
Anhedral with coronas
As in samples 17 to 12, some with coronas
Euhedral
Slightly resorbed
Smaller laths
Some larger crystals, no oscillatory zoned cores
Small laths with distinct preferred horizontal orientation
Small laths
Small laths



AUGITE
γ : c
NORM. FeO : MgO
TEXTURAL VARIATION
ORTHOPYROXENE
VOLUMETRIC MODE
x NORM. VALUES
10 20 30 40%
$2V_\alpha$
50° 55° 60° 65° 70° 75° 80° 85°
AVERAGE % FERROSILITE
15 25 35 45%
TEXTURAL VARIATION
ACCESSORY MINERALS
VOLUMETRIC MODES
x NORMATIVE VALUES
Some larger grains enveloping olivines
Clusters of interlocking grains enclosing small plagioclase laths
Rather small grains
Clusters associated with olivine, enveloped and penetrated by orthopyroxene
Somewhat larger grains
Anhedral prismatic crystals with distinct preferred horizontal orientation
Subophitic, some enclosing hypersthene prisms
Ophitically enclosing plagioclase and some olivines
Large crystals poikilitically holding highly resorbed olivines
Small anhedral grains
Rather small crystals
Some prismatic cores free from chadacrysts
Poikilitic oikocrysts optically continuous over large areas, enclosing all main constituents present.
Present also as small grains forming part of the augite clusters
Mostly with augitic intergrowths
Some cores free of chadacrysts
Prismatic crystals with indistinct preferred horizontal orientation
Subophitic with augitic intergrowths
Ophitically enclosing plagioclase and some olivine
Large crystals poikilitically holding highly resorbed olivines
Small anhedral grains
Alk. felspar
Quartz
Orthoclase
Quartz
Nepheline
Biotite
Hornblende
Magnetite
Ilmenite
1758′ 14
1528′ 13
1430′ 12
1398′ 11
1091′ 10
786′ 9
716′ 8
7
6
5
185′ 4
43′ 3
12′ 2
0′ 1
CENTRAL ZONE
BASAL ZONE

ON THE GRANITE-HORNSTONE CONTACT AT SLIPPERS BAY

By

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(Submitted January, 1957)

ABSTRACT

The interdigitated granite-hornstone contact zone at Slippers Bay is examined and described. A comparison is made of the criteria which could be interpreted as evidence of in situ formation of the granite with evidence indicating a liquid intrusive replacement. The chemistry, field relationships and petrography of the granite leave no doubt of the intrusive nature of the contact at Slippers Bay.

CONTENTS

I. INTRODUCTION

Owing to their bearing on the granite problem, the plutons of the south-western corner of the Cape Province, South Africa, have recently attracted the attention of several writers. Recent publications dealing with the problem are: a comprehensive description of the granites as a whole by D. L. Scholtz (March, 1946), a description of the celebrated granite-hornstone contact at Sea Point by F. Walker and M. Mathias (June, 1946), and an examination of the inclusions in the granite by C. Boocock (1951).

The crux of the controversy about the genesis of granite bodies is the part played by a granite magma. Some writers have discarded the idea of granite magma, claiming that it "is and always has been a hypothetical concept"; others maintain it as an essential; yet others advocate some middle way, claiming only limited amounts of magma. It was stressed by H. H. Read (1951) that generalisations in this sphere can serve no useful purpose and that evidence relating to the genesis of a specific granite body should come from a study of its field relationships. This evidence may then be reinforced by chemical and petrological findings, but not dominated by them. The relative importance of field and laboratory evidence is reversed by M. L. Keith and O. F. Tuttle (1954) because of conflicting interpretations of field evidence.

While a magmatic origin of the Cape granites is generally accepted, there are different opinions about the sequence of events attending the advent of the granite. The Sea Point contact is the only contact zone of the plutons which has been investigated in great detail. At the suggestion of Prof. D. L. Scholtz an investigation of another granite-hornstone contact at Slippers Bay, 80 miles to the north of Sea Point, was undertaken, the object of the investigation being an attempt to shed further light on some of the problems. Although this latter contact differs considerably in appearance from its Sea Point counterpart, observations there have led to general conclusions similar to those arrived at by workers at Sea Point and on the Cape granites as a whole.

II. HISTORICAL

Attention was first focussed on the Cape granites (Fig. 1) early last century when the Platteklip Gorge granite-hornstone contact was discovered by Captain B. Hall in 1813. The Sea Point contact was described by C. Abel in 1818, but it was not until Charles Darwin visited the locality in 1844 that an attempt was made to explain the phenomena present. The contact was subsequently described by E. H. L. Schwarz (1913) and A. R. E. Walker (1929). The following is a description of the contact at Sea Point based mainly on the account by F. Walker and M. Mathias, with a brief review of the opinions expressed.

At Sea Point, the contact between the granite and the sediments into which it is intrusive is not sharply defined, owing to the development of

a migmatite zone which contrasts strongly with both granite and sediments. Where it is free from inclusions, the granite is extremely uniform. It is composed of white to pale pink phenocrysts of microcline microperthite set in a matrix of quartz, microline microperthite, oligoclase and biotite. Cordierite and tourmaline occur scattered through the granite, while accessories include zircon, apatite, magnetite and pale pink almandine garnet. The latter is considered to be xenolithic in origin. Although lengths of up to 15 cm. have been recorded, the average length of the potash feldspar phenocrysts is 4 cm. Microscopically the phenocrysts are found to contain abundant inclusions of quartz (sometimes graphically intergrown), biotite and plagioclase. These inclusions tend to be arranged in a zonal manner, leaving the core relatively clear.

Veins of aplite are common in the vicinity of the contact, both in the granite and in the sediments. These aplites, attaining a maximum width of two feet, are found both along the planes of parting in the sediments and cutting across them; grain-size variations are proportional to the thickness of the vein. To the north-west of the contact, within the main body of sediments, a thicker mass of aplite-like rock is found. This mass shows both concordant and discordant relationships to the sediments and encloses three large xenoliths of hornstone. A striking feature here is the structural continuity between the xenoliths and the main body of sediments.

The sediments in contact with the granite belong to the pre-Devonian Malmesbury series, the oldest rocks exposed in the Western Province. Composed of alternating beds of pelitic and psammitic rocks, these sediments had already undergone extensive deformation before the advent of the granite. The major fold axes of the sediments trend approximately north-west, parallel to the direction of elongation of the main granite plutons. The deformation of the sediments has produced a well defined slaty cleavage or bedding fissibility in the more argillaceous members, and at Sea Point the contact zone is parallel to this direction. Contact metamorphism has altered the sediments to a dark, hard hornstone with conspicuous poikiloblasts of cordierite. Away from the contact, the metamorphic effects become less pronounced, but because of the concordance of the contact, the tracing of progressive metamorphism on a strike line is impossible.

The migmatite zone is some 70 yards wide and is characterised by an extremely wavy and streaky appearance, caused by the mingling of the hornstone and granite materials in a *lit par lit* fashion. The streakiness is thus aligned parallel to the cleavage in the main hornstone body, like the cleavages of the sediment in the mixed zone. As the main body of granite is approached, the amount of sedimentary material in the granite decreases and there is a gradation into the normal granite. It appears that the nature of the mixing has been selective, for some of the hornstone material is free from granitic substance, the unaffected material in general being more quartzose. The impression is gained that the more micaceous portions were impregnated by the granite more easily, possibly in consequence of their chemical similarity.

Microscopically the metamorphosed sedimentary facies of the migmatites is found to be a cordierite-biotite-hornfels with quartz and granular soda and potash feldspars. The biotite flakes show a more distinct parallelism than in the main hornstone body, while cordierite crystals have more definite outlines. The accessory minerals are similar to those in the main hornstone

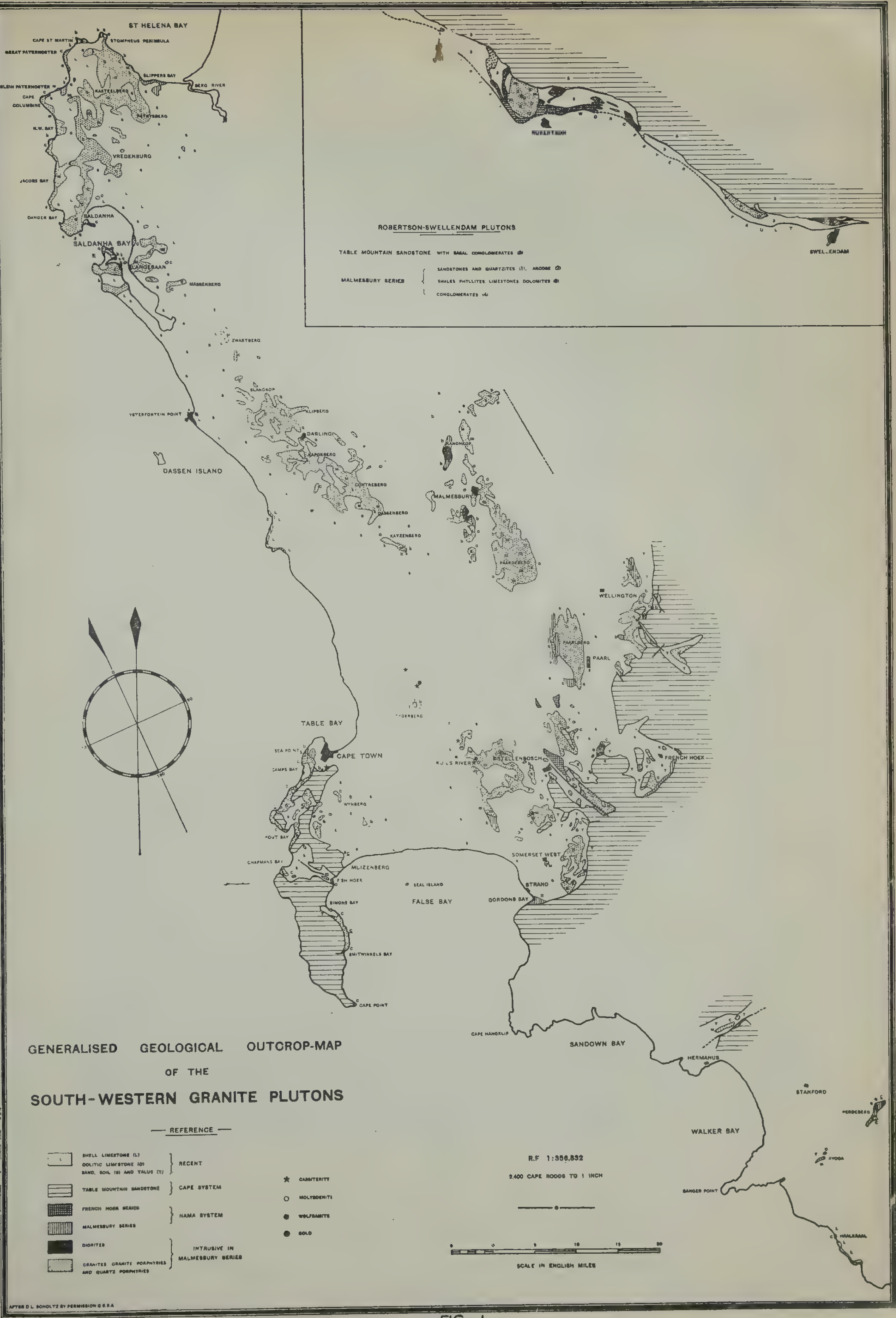

ROBERTSON-SWELLENDAM PLUTONS
TABLE MOUNTAIN SANDSTONE WITH BASAL CONGLOMERATES
MALMESBURY SERIES
SANDSTONES AND QUARTZITES (1), ARCOSE (2)
SHALES PHYLLITES LIMESTONES DOLOMITES (3)
CONGLOMERATES (4)
ROBERTSON
SWELLENDAM
ST HELENA BAY
CAPE ST MARTIN
STOMPNEUS PENINSULA
GREAT PATERNOSTER
SLIPPERS BAY
BERG RIVER
KLEIN PATERNOSTER
KASTEELBERG
CAPE COLUMBINE
PATRYSBERG
N.W. BAY
VREDENBURG
JACOBS BAY
DANGER BAY
SALDANHA
SALDANHA BAY
LANGEBAAN
MASSENBERG
ZWARTBERG
SLANGKOP
KLIPBERG
YSTERFONTEIN POINT
DARLING
DASSEN ISLAND
KAPOKBERG
CONTREBERG
MALMESBURY
KATZENBERG
PAARDEBERG
WELLINGTON
PAARLBERG
PAARL
TABLE BAY
CAPE TOWN
SEA POINT
CAMPS BAY
WYNBERG
STELLENBOSCH
FRENCH HOEK
HOUT BAY
CHAPMANS BAY
MUIZENBERG
FISH HOEK
SEAL ISLAND
SOMERSET WEST
STRAND
GORDONS BAY
FALSE BAY
SIMONS BAY
SMITWINKELS BAY
CAPE POINT
CAPE HANGKLIP
SANDOWN BAY
HERMANUS
STANFORD
WALKER BAY
DANGER POINT
GENERALISED GEOLOGICAL OUTCROP-MAP
OF THE
SOUTH-WESTERN GRANITE PLUTONS
REFERENCE
SHELL LIMESTONE (L)
OOLITIC LIMESTONE (O)
SAND, SOIL (S) AND TALUS (T)
RECENT
TABLE MOUNTAIN SANDSTONE
CAPE SYSTEM
FRENCH HOEK SERIES
MALMESBURY SERIES
NAMA SYSTEM
DIORITES
GRANITES GRANITE PORPHYRIES AND QUARTZ PORPHYRIES
INTRUSIVE IN MALMESBURY SERIES
CASSITERITE
MOLYBDENITE
WOLFRAMITE
GOLD
R.F. 1:356,832
2,400 CAPE ROODS TO 1 INCH
SCALE IN ENGLISH MILES
AFTER D.L. SCHOLTZ BY PERMISSION G.S.S.A

FIG. 1

body. The granitic material of the mixed zone differs quite widely from the normal porphyritic granite. The average grain-size is between 0.5 and 1.0 mm. Biotite is less common; muscovite more so. Plagioclase, having a higher Ab content, is present in larger proportions. Smaller streaks of granite are more aplitic, while the larger ones tend to be similar texturally to the porphyritic granite. Possibly the most remarkable feature in the migmatite zone is the development of large crystals of potash feldspar, up to 6 cm. in length, in portions of the sedimentary fraction of the mixed zone. These large crystals are of random orientation and appear to have formed just as readily in the sedimentary portions of the mixed zone as in the granite, without disturbing the schistosity of the former. Microscopically they are similar to the phenocrysts in the normal granite and do not have as inclusions material characteristic of the sediments. F. Walker and M. Mathias believe that these large feldspars in both the granite and sediments originated as a result of late potash enrichment by circulating solutions. C. Boocock, however, points out that in the sediments the crystals are surrounded by rims of granitic material which trail off into the surrounding rock. This, he suggests, is indicative of the large feldspars having crystallised from granitic material injected into the sediments. A careful examination of the migmatite zone at Sea Point reveals that these crystals of potash feldspar are *only found* in those portions of the hornstone which show signs of contamination by granitic material. The large feldspars are furthermore totally lacking in the hornstone beds once the contact has been left behind.

Inclusions of Malmesbury sediments in the granite are ubiquitous, although they decrease in number away from the contact. Near the contact inclusions up to 200 feet in length are present; composed of hornfels, they show local migmatisation with an erratic distribution of potash feldspar phenocrysts. Smaller inclusions are, however, more common and, like the larger ones, show evidence of movement and plastic deformation. D. L. Scholtz believes that local concentrations of some of these smaller inclusions are due to the disruption of larger xenolithic bodies. Further from the contact the inclusions appear to be much more mixed with granitic material and exhibit a granitoid texture; their grain-size is somewhat less than that of the surrounding rock, while biotite is more plentiful. According to C. Boocock, the inclusions in the granite can be divided into two groups, xenoliths of hornfels and granitic xenoliths, the latter representing an advanced stage of alteration of the former.

All workers at Sea Point and other parts of the Cape granites have agreed in principle that the granite had intruded in a molten state. The reconstruction of events at Sea Point by Charles Darwin, long ago, merits special note, especially as it is based entirely on field evidence and a keen and critical sense of observation: "As we leave the junction (between the granite and clay slate), thin beds and, lastly, mere films of the clay slate are seen, quite isolated, as if floating, in the coarsely crystalline granite; but, although detached, they all retain traces of the uniform N.W.-S.E. cleavage. This fact has been observed in other similar cases and has been advanced by some eminent geologists as a great difficulty on the ordinary theory of granite being injected whilst liquified, but if we reflect on the probable state of the lower surface of a laminated mass like clay slate, after having been violently arched by a body of molten granite, we may conclude

that it would be full of fissures parallel to the planes of cleavage; and that these would be filled with granite, so that wherever the fissures were close together, mere parting layers or wedges of slate would depend into the granite. Should, therefore, the whole of the rock become worn down and denuded, the lower ends of these dependent masses or wedges would be left quite isolated in the granite; yet they would retain their proper lines of cleavage, from having been united whilst the granite was fluid, with a continuous covering of clay slate."

E. H. L. Schwarz (1913) interpreted the migmatic phenomena as evidence of replacement, and claimed that the orientated xenoliths were island remnants in a "pseudomorph" of slate in granite. In criticising the interpretations of E. H. L. Schwarz, D. P. McDonald (1913) pointed out that Daly's "Overhead Stoping" theory would account satisfactorily for the observed phenomena at Sea Point. F. Walker and M. Mathias favour a magma intrusion with subsequent potash enrichment to account for the phenocrysts. C. Boocock's criticism of this view has already been noted. D. L. Scholtz believes that plutons of two distinct ages were emplaced with the magma rising along the cores of the unsymmetrical anticlines in the surrounding sediments, aided by the process of overhead stoping, thus partly accounting for the great prevalence of concordant contacts and the general abundance of xenoliths of very variable size. Compositional variations in the masses are believed to be the result of gravitational differentiation and contamination.

The study of a large feldspar phenocryst from the Cape granite at Sea Point by S. J. Shand (1949) led him to conclude that the crystal could only have grown in a liquid environment. This feldspar measures 58 x 31 mm. and consists of a microcline core 47 x 19 mm., surrounded by a rim of plagioclase. A few idiomorphic inclusions of plagioclase are present in the core but are so small in relation to the microcline that it seems certain that the microcline began to grow first. On the edges of the microcline are thin films of quartz which are followed by the plagioclase mantle. In this latter mantle grains of biotite and quartz, which increase in size outwards, are included. Microcline is not present as an inclusion in the mantle. The matrix enclosing the large crystal is composed of quartz, microcline (devoid of plagioclase mantles), plagioclase and biotite. The phenocryst must, therefore, have begun to grow in a potassic liquid and subsequently migrated to a more sodic environment in which biotite and quartz were crystallising. It is further stated: "It is difficult to believe that aqueous solutions or ions migrating through solid rock could have formed a heavy mantle of plagioclase on the large microcline and no mantles at all on smaller crystals of the same mineral, if both had been present at the same time."

III. THE CONTACT ZONE AT SLIPPERS BAY

General.

Some 80 miles N.N.W. of the Sea Point contact, another contact between the Cape granites and the Malmesbury sediments is exposed on the seashore. The exposure occurs in a small cove, known locally as Slippers Bay, which lies on the southern side of St. Helena Bay between Stompneus and Velddrif. Surface deposits extend almost to the high water mark, but below this the

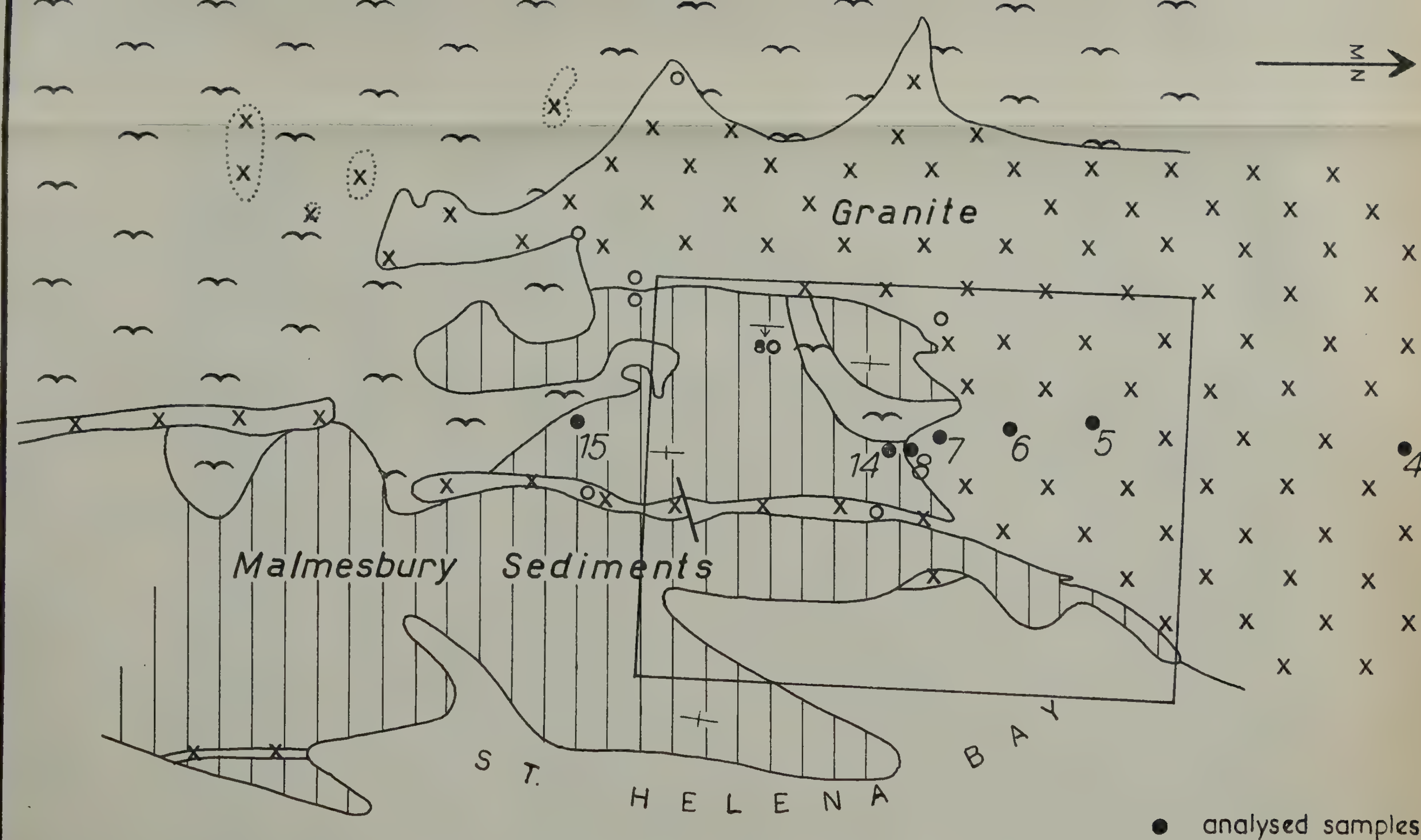
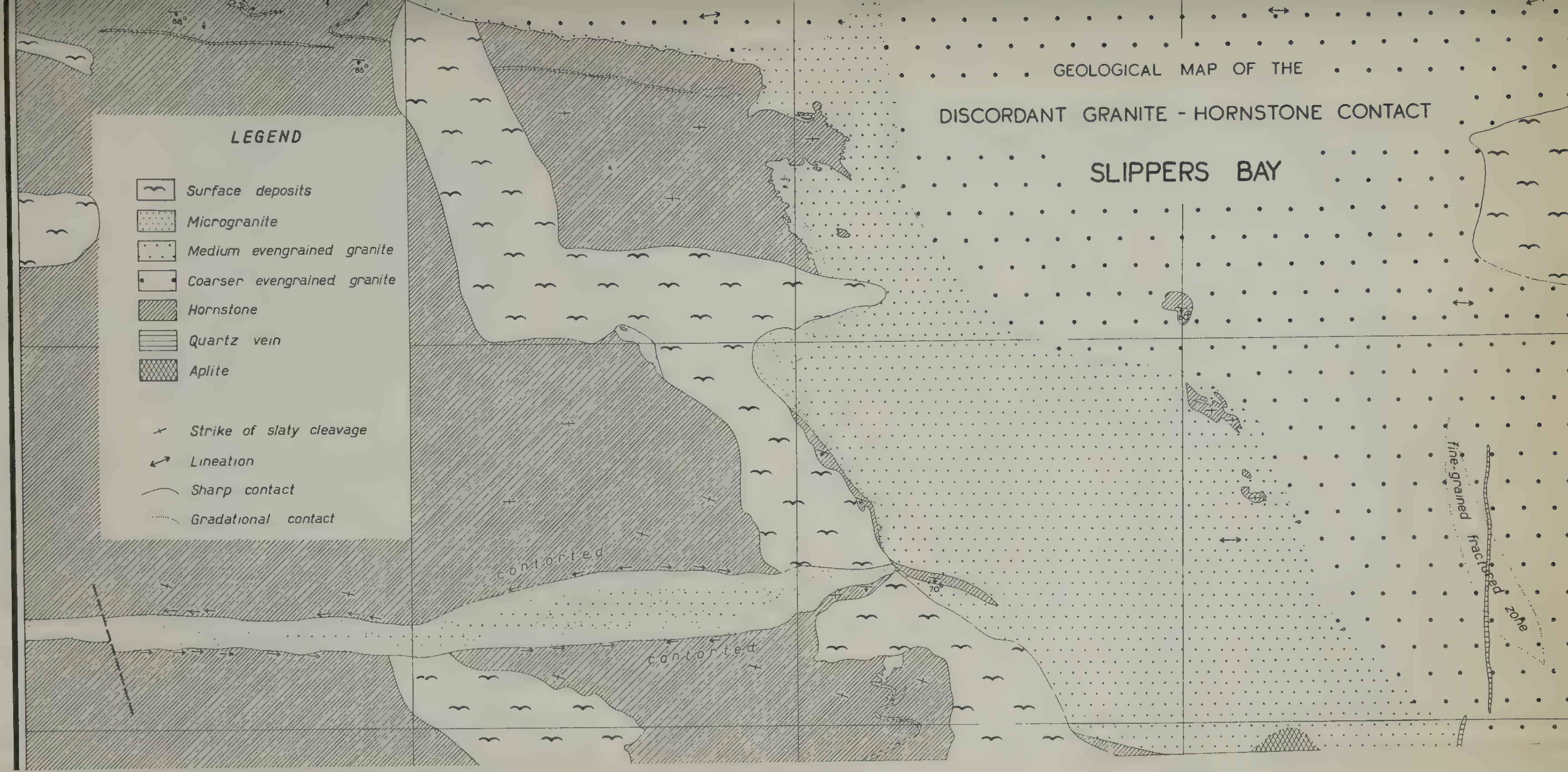


FIG 2. MAP SHOWING LOCALITY OF SOME OF THE SAMPLES TAKEN AT SLIPPERS BAY


GEOLOGICAL MAP OF THE
DISCORDANT GRANITE - HORNSTONE CONTACT
SLIPPERS BAY
LEGEND
Surface deposits
Microgranite
Medium evengrained granite
Coarser evengrained granite
Hornstone
Quartz vein
Aplite
Strike of slaty cleavage
Lineation
Sharp contact
Gradational contact
contorted
contorted
fine-grained fractured zone

coast has a rocky fringe and it is in this that the contact may be observed. For the most part of the contact is extraordinarily sharp and of a perfectly concordant nature, being again parallel to the cleavage of the sediments which strike north-south in this area. Along the concordant contact the granite exhibits a convex curvature towards the hornstone which can be cleanly scaled away from the granite by light blows with a hammer (Plate I, ph. 1). This concordant contact can be traced intermittently along the shore for about half a mile and as the granite is on the landward side, only a narrow strip of sediment is exposed in the intertidal area. In the Slippers Bay cove the granite was found to cut abruptly across the cleavage planes of the sediments, giving rise to a local discordant contact some 9 feet in length. This was the feature which focussed attention on the area (Plate I, ph. 2). This area was first mapped on a scale of 1:360 by means of a plane table and telescopic alidade (see Fig. 2). The discordant contact zone was then covered by a 25′ x 25′ grid, each portion of which was divided into squares 5′ x 5′. For the areas requiring more detail a 1′ x 1′ grid was drawn on the outcrops with chalk. A map of scale 1:24 was then constructed on square paper (see Fig. 3). Sampling presented a somewhat arduous task owing to the smoothness of the seawashed outcrops. Emphasis was placed on obtaining representative samples of the different facies of the granite in a north-south direction across the discordant part of the contact. The samples were orientated in the field with reference to a horizontal plane and the magnetic meridian. As can be seen from Figs. 2 and 3 and Plate I, ph. 2, the junction between the concordant and discordant parts of the contact is abrupt on the west, but to the east there is a gradual transition between the two as the contact swings out to sea. Where the granite is in contact with the extremities of the cleavages in the sediment, the junction between the two rock types is ragged and interdigitated. An examination of Plate I, ph. 4 shows how some of the outermost extremities of the bedding planes in the hornstone seem to have been forced apart and slightly bent, as if the granite material selectively exploited weaker beds and pushed aside the more resistant cusps during emplacement. Despite this interfingering between granite and hornstone, there is always a clear cut contact between them and there is no trace of migmatisation as at Sea Point. As the selective bending of the prongs of sedimentary material in the granite could not have been accomplished in the solid state, convincing evidence is provided of the liquid condition of the granite at the time of intrusion.

In the immediate vicinity of the discordant section of the contact, three distinct textural types of granite with seriate fabric relationships can be distinguished. Where the granite is in contact with the extremities of the cleavage and bedding planes in the sediment, it is in the form of a microgranite with an average grain-size of 0.68 mm. In this microgranite are pockets of slightly coarser material which possibly represented volatile rich patches during solidification. The fine, even-grained microgranite facies is limited to the granite in contact with the extremities of the cleavage planes in the sediment, and its width does not appear to exceed eight inches. North of the discordant contact, the microgranite *grades* into a medium even-grained facies with average grain-size of 1.37 mm.; this facies attains a maximum width of about five feet north of the microgranite. This medium even-grained granite forms a continuous band beyond the microgranite and swings round to follow the concordant contact where it is in contact with

the sediments. It narrows rapidly, however, and its place is taken by a coarser-grained granite with average grain-size 1.97 mm., into which the medium-grained granite grades both north of the discordant section and west of the concordant contact (see Fig. 4). Lineation in the granite is limited to the vicinity of the concordant part of the contact and is best displayed by the orientation of the quartz grains parallel to the contact.

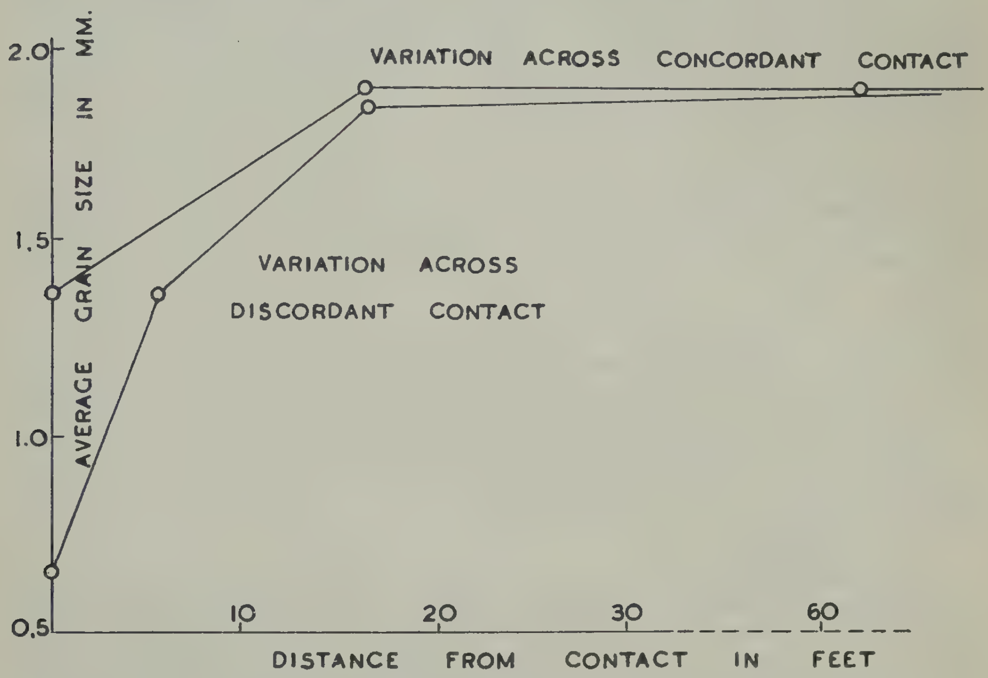


Fig. 4. Grain-size variations of granite across discordant and concordant sections of the contact.

Farther north, some 2,500 feet from the discordant contact, the granite last mentioned is in turn in sharp contact with a coarse markedly porphyritic granite, which is intrusive into the even-grained granite. The two are separated by a finer-grained facies about eight inches wide, which is present along the entire exposed contact.

Where exposed in the area, the Malmesbury sediments are represented by black semi-pelitic biotite-rich hornfels. Slaty cleavage is strongly developed, parallel to the bedding, dipping vertically or nearly so. The large cordierite poikiloblasts, so conspicuous at Sea Point, are not in evidence, consequently the rock is not maculose and presents an even-weathered surface. Quartz veins with eye-like swellings are numerous, occurring both along the partings in the hornfels and cutting across them, exhibiting ptygmatic folding. In size these veins range from microscopic stringers to veins half an inch or more in width. Granite veins in the hornfels vary in size from the fine-grained type, where interfingering has taken place along the discordant portion of the

contact, to aplite-like bodies up to six inches in width and to the dyke-like body which reaches a maximum width of three feet in the hornfels. Along the contact of this dyke, the quartz stringers and cleavage planes of the sediments show intense horizontal drag deformation. As at Sea Point, the grain-size of the vein material varies proportionately with the width of the body. In the dyke there is an increase in grain-size from the margin to the centre. Ptygmatic folding is exhibited by several of the smaller veins of granitic material in the sediments (Plate I, ph. 5). Veins which are parallel to the planes of flow cleavage in the sediments tend to be straight, the contortions, where present, being confined to the portions of the veins which cut across the bedding in the sediments. A tentative explanation of these ptygmatic structures might be sought in a local internal movement of the sediment by creep or flow folding parallel to the parting planes before complete solidification of the vein material. The writer believes that such movement would deform the granite material cutting across the cleavages, yet retain the linear nature of veins parallel to them. The majority of these veins can be traced back into the main body of granite via the discordant contact, but it is interesting to note the behaviour of the small veins which enter the hornfels from the conformable contact (see Fig. 3). Cutting across the strike of the cleavage for a short distance, they swing round to parallel the cleavage directions. At its source on the conformable contact, the longest of these veins is about one inch wide; narrowing rapidly, it pinches out fifteen feet from where it started. Microscopically the contact between the vein material and the sediment is sharp, being delineated by a line of biotite flakes belonging to the latter. Small schlieren of hornfelsic material with their long axes parallel to the length of the vein are shown in Plate II, ph. 4, while the partial scaling off of another fragment is shown in Plate II, pm. 1.

Inclusions of the hornfels in the granite exhibit similar structural phenomena to those displayed in the Sea Point migmatite zone. Immediately north of the discordant contact, small islands of hornfels occur in the finer-grained facies of the granite. Their structural continuity with the country rock is apparent in Plate I, ph. 3. Twenty-five feet farther north, two more islands of sedimentary material are found, the larger measuring 18 x 25 inches. Like the smaller inclusions, they too are in structural continuity with the country rock and show no signs of movement. In all respects they are similar to the wall-rock hornfels and display sharp contacts and no admixture of granite material. Several hundred yards westward from the contact, outcrops of coarser non-porphyritic granite appear again from beneath the surface cover. Inclusions in this coarser granite are in strong contrast to those in the fine-grained granite; having rounded edges, they show signs of intensive hybridisation by the granitic material. In the example shown in Plate I, ph. 6, it can be seen that a granitoid texture is developed in the inclusion. The size and shape of these inclusions vary from fairly well rounded clots of about six inches across, to a large, irregular body 600 x 150 feet, first recognised by D. L. Scholtz in the hills above the contact, two-thirds of a mile to the west. Despite its pronounced granitoid texture, it is in strong contrast to the enveloping granite, being finer-grained and darker in colour. Mineralogically this hybrid rock is markedly granodioritic (see Table I). Its edges are smoothly rounded, contact with the granite is sharp, and in places it can be seen to be veined by the surrounding granite.

Petrography.

(a) *Granite.*

In hand specimen the finer-grained granites are white to cream coloured; biotite is not abundant, occurring throughout the rock in small flakes. With an increase in grain-size, biotite increases in amount and the rock becomes mottled owing to the pink colouration of the potash feldspars. Despite textural variations from coarsely porphyritic to micrigranitic, the granite is mineralogically uniform, being composed of microcline microperthite, quartz, plagioclase and biotite in order of decreasing abundance.

In a paper submitted to the Department of Geology, University of Stellenbosch in January, 1956, J. Otto has described a scheme which facilitates and simplifies the field classification of texturally different igneous rocks. This scheme is made use of in Fig. 5, which is a schematic figure illustrating the interrelationships of the different textural types of granite in the vicinity of the contact.

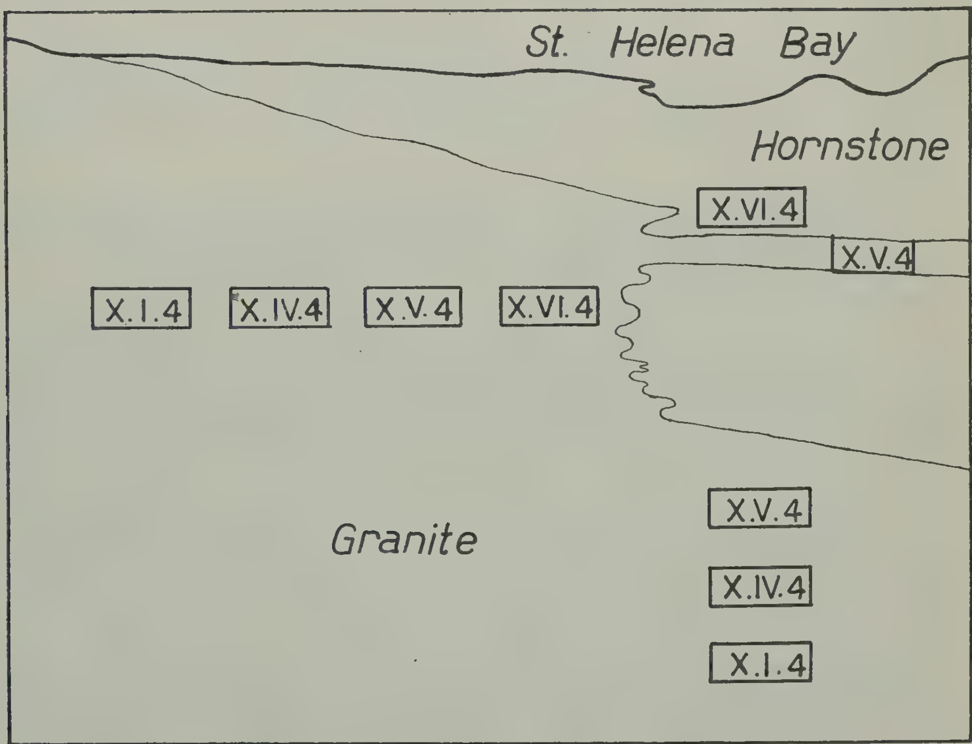


Fig. 5. Diagrammatic figure showing relationship of the different textural types of granite in the vicinity of the contact.

Where the grain-size of the rocks permitted, modal determinations were carried out with an integrating stage on three thin sections cut at right angles to one another from each sample. Measurements along 1 x 1 cm. grid-lines on three smoothed surfaces at right angles to one another on a hand specimen were employed for determining the volumetric percentage of the constituent minerals in the coarse types. This procedure was facilitated by staining the smoothed faces with sodium cobalti-nitrite (F. Chayes, 1952). To achieve the staining, the rock face was first swabbed with HF and left for five minutes. After being washed in running water, the area treated was sprayed with a solution of 4 gm. sodium cobalt-nitrite in 6 cc. H_2O. After another five minutes, the flat surface was again washed and swabbed clean. This provided a highly satisfactory manner of distinguishing macroscopically between the potash and plagioclase feldspars.

TABLE I — MODAL PROPORTIONS

	A	B	C	D	E	F	G
Quartz	38.00	33.70	36.40	39.10	36.20	34.00	62.50
Microcline Microperthite	39.80	48.00	46.00	43.70	41.30	23.10	—
Plagioclase	16.20	17.30	15.60	16.30	20.20	35.30	present
Cordierite	—	—	—	—	—	—	2.20
Biotite	0.80	1.00	2.00	0.90	3.10	4.40	35.30
Muscovite	—	—	—	—	—	3.20	—
Garnet	—	—	—	—	—	—	present
Apatite	—	—	—	—	—	—	present
Matrix and accessory Constituents	5.20	—	—	—	—	—	—

A. Contact microgranite.
B. Granite 7 feet north of contact.
C. Granite 20 feet north of contact.
D. Granite 1,200 feet north of contact.
E. Granite 1,000 feet west of contact.
F. Hybrid granodioritic rock.
G. Hornfels from contact.

Quartz. In thin sections cut parallel to the lineation where visible, the quartz grains were found to make up streaky sutured mosaics, the compotent grains of which were all characterised by undulose extinction. Away from the contact in the coarser-grained granites, the quartz forms normal ragged anhedrons with no apparent preferred orientation. Undulose extinction is the rule rather than the exception.

Strings of dust-like inclusions form curved trails in all quartz grains in the granites, a feature which is much less pronounced in the hybrid xenolith, where the grains are frequently devoid of any inclusions. Quartz appears intergrown with microcline microperthite in all facies of the granite except the microgranite, where this relation was not observed. The grain-size of the quartz varies from 0.3 mm. in the contact microgranite to 5 mm. in the porphyritic granite. A petrofabric diagram based on sections cut in an east-west direction across the lineation in the granite of the conformable contact shows a definite axial symmetry with the *c*-crystallographic axes of the quartz grains lying horizontally parallel to the lineation. This preferred orientation is not nearly as conspicuous, but is nevertheless present in sections cut east-west across the strike of the slaty cleavage in the hornfels a few feet from the granite sample referred to above (see Fig. 6a and 6b).

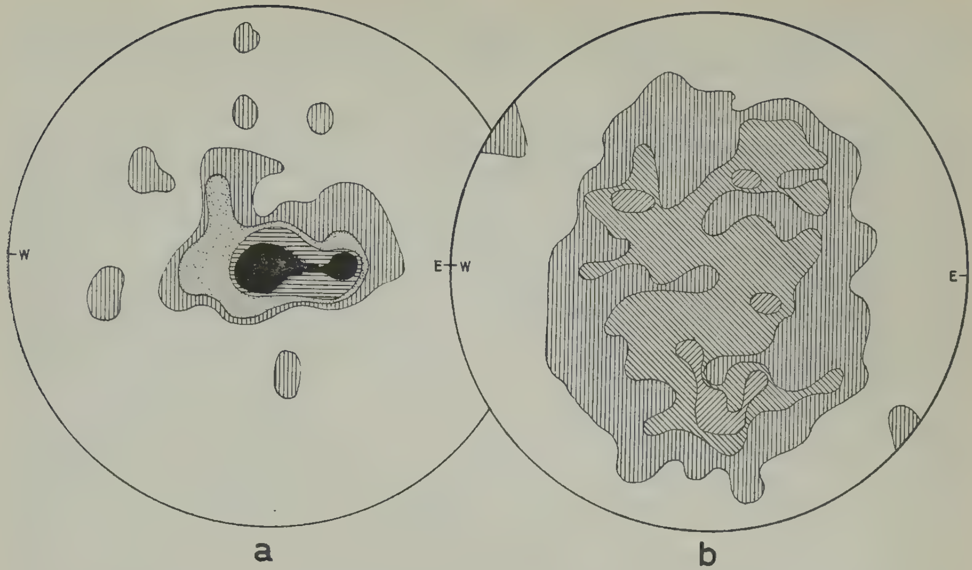


Fig. 6. Petrofabric diagram illustrating the orientation of the *c*-axes of 300 grains of quartz in granite (a) and 300 grains of quartz in the hornstone (b) on either side of the concordant contact. 0 — 1% vertical ruling, 1 — 2% right diagonal ruling, 2 — 3% left diagonal ruling, 5% stippled, 10% horizontal ruling and 13% black.

Potash Feldspar. This mineral consistently comprises the principal constituent of the granite and invariably appears as microcline microperthite. The value of the optic axial angle ranges from $2V_{\alpha} = 74°$ to $2V_{\gamma} = 85°$, variations being encountered within a single crystal. C. T. Potgieter (1950) records a similar variation in the granites of the George area and attributes the variation in the size of the optic axial angle to strain. Orthoclase was not encountered although measurements on the universal stage were carried out to test for its presence. The grain-size of the microcline microperthite varies from 0.5 mm. in the microgranite to 1.25 cm. in the coarse even-textured granite, while in the porphyritic granite phenocrysts of up to 3.75 cm. are common. In the hybrid rock occasional euhedrons of microperthite and plagioclase up to 9 x 4 mm. give the rock a micro-porphyritic habit.

The characteristic gridiron twinning is frequently present, while simple Carlsbad twins are not uncommon. Crystals devoid of twinning are, however, present in all facies of the granite and the hybrid xenolith.

Perthitisation appears to increase in intensity with the increase in grain-size. Untwinned crystals have more perthitic intergrowths than those exhibiting gridiron twinning. In the microgranite the perthite structures are confined to small string and stringlet types (H. L. Alling, 1932). These obtain their maximum development in the centres of their hosts, leaving the rims comparatively free of intergrowths. Intergrowths in the coarser granites become more irregular but still belong to the string type. They are not only wavy but may also bifurcate and in certain cases form fine marginal segregations

along the edges of their hosts. All perthitic strings do not show apathy towards quartz inclusions in the feldspar; frequently they are seen to be sharply truncated by the quartz or else to form indistinct rims about the inclusions. On the other hand areas of microcline microperthite or quartz infilled embayments are often devoid of perthitic intergrowths. None of the larger patch and plume types of perthite, which H. L. Alling believes to be indicative of replacement, was encountered in the granite. In the hybrid xenolith, however, the perthite is ragged and marginal blebs of plagioclase can be seen to vein their hosts. An examination of the crystallographic directions followed by the perthite stringers showed that $(15\,0\,\bar{2})$, (110) and (100) were favoured in that order of abundance.

Micrometric counts carried out with an integrating stage showed that up to 30% of a perthite crystal can be composed of plagioclase. From the chemical analyses of the feldspar the Or content of the potash feldspars from the porphyritic and even-grained granites was calculated as 66% and 56% respectively. The calculations revealed a deficiency in alumina and an excess of silica. By substituting silica for alumina in the calculation, the deficiency and the excess are cancelled out. The slight excess of silica then remaining is within the margin of accuracy of the analysis. According to the work of E. Spencer (1938) and O. F. Tuttle (1952), the temperature of formation of the microcline microperthite from both porphyritic and even-grained granites falls in the vicinity of 660°C. F. Chayes (1952) has noted the affinity of microcline microperthite to strained quartz and believes that pressure was intrumental in aiding unmixing of the feldspars.

Plagioclase. The plagioclase content of the rocks varies from 16% to 20%. The grains are euhedral and subhedral, and are never large enough to be conspicuous. Saussuritisation has been extensive, more so in the even-grained granites than in the porphyritic facies. The cores of the crystals are often most affected and in many cases the alteration has caused the obliteration of twinning in the affected part of the mineral. The twinning laws which are most abundant are Albite, Albite Ala and Carlsbad, all having evenly developed composition faces despite faint zoning.

Compositional determinations carried out with the universal stage showed that the anorthite content of the twinned plagioclase in the even-grained granites was between 13% and 19%. Where determinable, it was found that the zoned members had slightly more sodic rims, the increase being 3% or less. No progressive change in the lime content of the plagioclase was recognisable away from any contact. In thin sections cut from the porphyritic granite most of the twinned plagioclase returned a value of 23-25% An, zoned crystals again having a slightly more sodic rim. The above plagioclases are all of the low temperature variety.

The composition of the plagioclase feldspars was checked by the method described by R. W. Foster (1955), entailing the fusion of fragments and subsequent R.I. determinations of the glass. Selection of material was again facilitated by staining with sodium cobalti-nitrite. It was found that the plagioclase from the even-grained granite had an average composition of about 10% An ($n = 1.489$) and that from the porphyritic granite approximately 26% ($n = 1.511$-1.516). All liquids used were immediately checked on the Abbe refractometer. The difference in composition returned by the two methods

for the plagioclase from the even-grained granite might be sought in its more decomposed nature, which would lead to partial decalcification during the staining.

Untwinned or partially twinned plagioclase, while absent from the fine-grained granites, is present in small amounts in the coarser even-grained granites and porhyritic types. These grains exhibit very marked zoning of the progressive type (R. C. Emmons and V. Mann, 1953), and often show twinned cores grading into untwinned rims. The microscopic investigation of these feldspars in thin section showed that the composition of the cores varied from An_{30} to An_{50}, while the rims consisted of oligoclase. Plate II, pm. 2 depicts such a crystal traversed by a fracture filled with rim material and intergrown with another plagioclase. D. L. Scholtz records that myrmekitic intergrowths in the acid mantles of the plagioclase form a common feature of the granites.

In the hybrid rock plagioclase, whether twinned, partially twinned or untwinned, is the most abundant mineral present, zoning being again very pronounced in all types. The composition of the cores is analogous (or nearly so) to the last type described, while the rims too are oligoclase. Quartz intergrowths in the untwinned mantles are common and characteristic.

Biotite is the only mafic mineral of any importance in the granite, but it is present in small quantities and rarely constitutes 3% by volume of the porphyritic granite. Chloritisation, especially in the even-grained granites, is in places far advanced, all stages of alteration from brown-green biotite to chlorite being present. Included zircon grains give rise to the characteristic haloes. In hand specimen of the hybrid rock, xenolithic biotite appears to be more abundant but modal analyses indicate that only 4% is present.

Muscovite. Primary muscovite is not present in any of the granites under consideration. When present, it is in the form of fine sericite flakes, apparently derived from the alteration of feldspars. In the hybrid inclusion aggregates of this mineral are usually associated with pinite. D. L. Scholtz regards it as an alteration product of cordierite. C. Boocock, in the case of the granites of the Cape Peninsula states, "The various degrees of alteration of the cordierite have resulted from the metasomatic introduction of variable amounts of potash-rich alkaline material from the magma. The final product of alteration appears to be muscovite with pinite, representing an incomplete stage. The cordierite was originally formed largely from the sericite, chlorite and to a small extent from the iron in the sediments. Under the influence of the potash-rich alkaline solutions, this process is reversed."

(b) *Malmesbury Sediments.*

Thin sections of the strongly cleaved hornfels show a preponderance of biotite flakes orientated parallel to the cleavage traces. Quartz, which is also abundant, is in the form of irregular, sometimes elliptical grains. Parallelism in a direction parallel to the cleavage is present but not marked. Plagioclase, An_{15}, cordierite, apatite and almandine garnet are present in lesser amounts, the rounded outlines of some of the latter indicating an allogenic origin. The metamorphism of these sediments has already been described

in detail by D. L. Scholtz, F. Walker and M. Mathias and C. Boocock, the course of events being similar to the normal thermal and regional metamorphism of the Malmesbury sediments.

(c) *Heavy Mineral Investigations.*

The heavy concentrates were extracted from four samples of granite taken on a north-south line across the interdigitated portion of the contact. It was thus possible to examine and compare the heavy minerals from the contact granite, granite 75 feet and 1,200 feet north of the contact, and from the porphyritic granite with those of the sediments. From the granites five or six pounds of material was crushed in an iron mortar to -90 mesh, care being taken to avoid grinding. The crushed material was then panned and concentrated further by treatment with bromoform and methelene iodide. The concentrates were then cleared by boiling in dilute HCl. It was felt that a single sample taken from the hornfels might not be sufficiently representative; accordingly fifteen samples were taken from various parts of the exposure. From these little blocks of one cubic inch were cut, crushed and thoroughly mixed before proceeding with the investigation.

The heavy mineral assemblage obtained from the granite comprises magnetite, zircon, pyrite, bluish-green tourmaline, epidote, fluorite and occasionally pale pink almandine garnet, which is more abundant in the porphyritic granite. Epidote is more abundant in the sample taken 1,200 feet north of the contact, where the granite is redder than usual. The colour is attributed to hydrothermal alteration. The hornfels yielded a similar crop of heavy minerals in which garnet was more abundant and zircon less so than in the granites. If the magnetite in the granite concentrates is disregarded, zircon becomes the most abundant mineral present.

The concentrates were mounted in Canada balsam and with the help of a mechanical stage and graduated eyepiece, length and breadth measurements were carried out on 100 zircons from each sample. Further, by using the universal stage, it was possible to study the shape of the grains.

Of the zircons from the granites, 97% are idiomorphic with smooth well-developed prisms and bipyramids, the basal pinacoid being occasionally present (Plates II and III, grp. B and C). The remaining 3%, which show signs of rounding, were derived from the porphyritic granite. The size of the grains varies between 0.50 x 0.23 mm. and 0.06 x 0.03 mm. There is little change in the colour of the zircons; 90% of the grains are colourless and relatively clear, with some with pink or yellow tints and cloudy grains constituting the remaining 10%. Zoning is a prominent feature of the zircons, 92% of which exhibit this phenomenon. The zoning is largely confined to the outer parts of the crystal, although many grains are zoned throughout, the zoning following the crystal outlines. The remaining 8% consists of a number of clear, glassy grains and opaque crystals in which the zoning may or may not be present. Inclusions in the zircons are common; indeed, a grain without some type of inclusion is rare. The inclusions range in size from dust-like specks to dark or translucent needle- and rod-like bodies up to 0.07 mm. in length. In addition 12% of the granite zircons contain definite cores which are elongate, rounded and in optical continuity with their hosts. The sizes of the cores vary between 0.03 x 0.02 mm. and

0.14 x 0.06 mm., there being no apparent relation between the size of the core and that of the host. The elongation index of the zircons is larger than 1 : 2, although many are borderline cases. Despite this relatively constant length to breadth ratio, there is a gradual but marked increase in the grain-size away from the contact (see Fig. 7). If the rounded grains are disregarded, three types can be distinguished from among the zircons of the granite. They are in order of decreasing abundance.

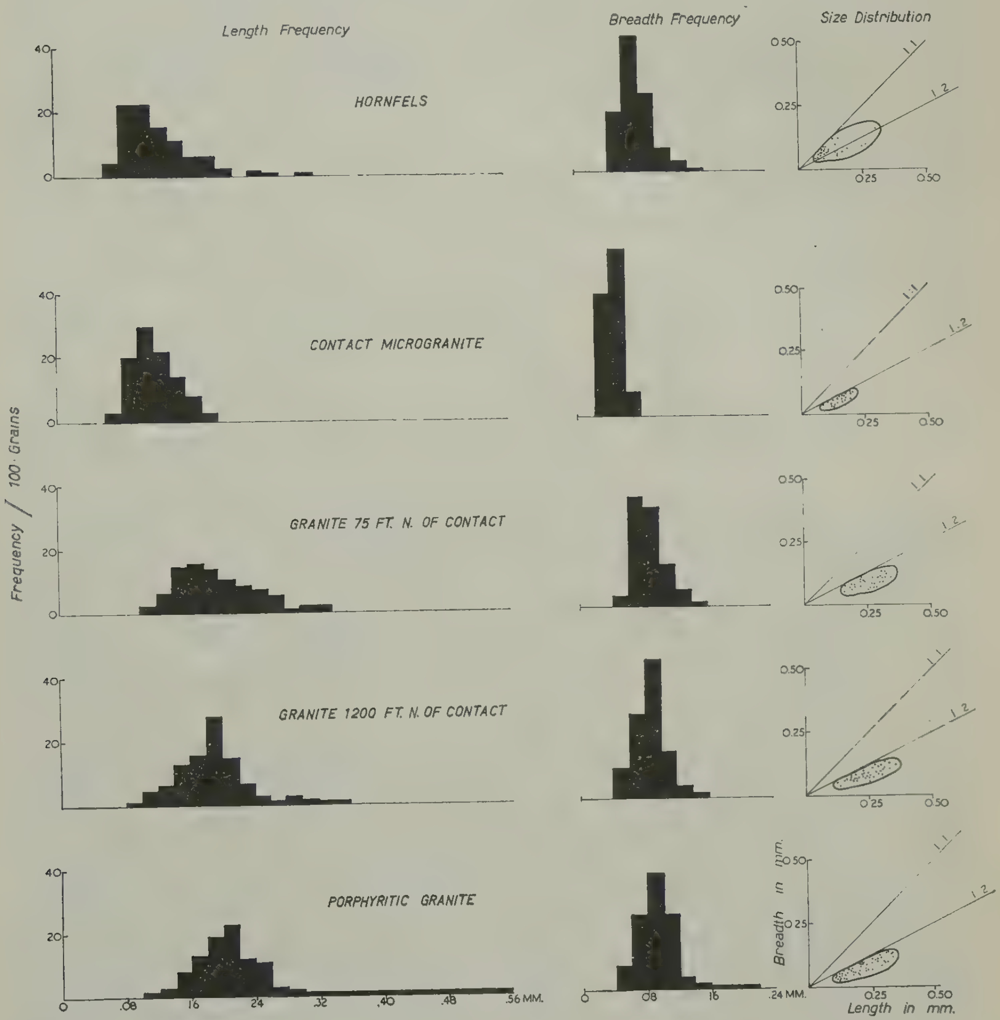


FIG. 7

Zircon histograms etc.

1. Stumpy zoned grains, sometimes semi-opaque with or without cores.
2. Elongated grains, zoned, clear with or without cores.
3. Clear, glassy unzoned crystals, occurring only in the porphyritic granite.

Of the zircons derived from the hornfels, 95% can be classified as rounded or well rounded, the remainder as subangular. The zircons have rough, pitted exteriors, owing to the abrasion they have undergone (Plate III, grp. A). Among the zircons from the hornfels are grains up to three times the size of the largest grains from the contact microgranite. Angular zoning in the grains is similar to that in the granite, 76% of the grains being thus affected. The nature of the inclusions, too, differs in no way from what has been described before, but only 4% are cored. Elongation index ratios are for the most part between 1 : 1 and 1 : 2 (see Fig. 7). As many of these rounded grains have dimensions of less than 0.1 mm., it would appear that the aeolian agencies supplemented aqueous agencies in the transport of the original sediment. The salient characteristics of the zircons derived from the granite and adjacent hornstone may be summarised as follows:

	Shape	Exterior	Zoned	Cored	Fractured	Colour
Granite	Idiomorphic	Smooth with flat crystal faces.	Angular 92%	12%	31% heavily fractured, often radially	Colourless 90%
Hornstone	Rounded	Rough and pitted	Angular 76%	4%	Free from fracture	Colourless 91%

Following A. F. Wilson (1950), the present writer examined zircons from both granite and hornfels in ultra-violet light. The examination was carried out in a photographic dark room on the stage of a binocular microscope. Both groups of zircon, however, gave no indication of fluorescence. +0.1 milligrams of the zircons were then handpicked and examined spectrographically. As could perhaps be expected, no difference in chemical composition could be determined. Among the lines identified were Zr, Si, Al, Mn, Hf and Ca. The remaining minerals in the heavy concentrates from both granite and hornstone are, generally speaking, similar in all respects.

Zircon has been regarded as being stable enough to withstand advanced metamorphism and has been used as a key to rock history. There is, however, a definite limit to this stability. J. A. Butterfield (1936), F. Smithson (1937) and G. Bond (1948) have described outgrowths on zircon in relatively unaltered sediments, while A. Poldervaart and J. W. von Backström (1950) showed that in the Kaaien granulitic rocks at Kakamas, zircon crystals have partially grown together. This feature they attribute to granulation and recrystallisation in a metamorphic grade beyond the sillimanite zone. L. T. Aldrich et alia (1955) record that zircons from the Cape granites showed annealing of crystal structures after roasting for five minutes at 450°C. M. Wyatt (1954) examined zircons from a Scottish granite, the surrounding Moine granulites, granite contaminated by granulite and altered, granitised sediments from the vicinity of the contact. In the granite proper, the zircon was idiomorphic and commonly zoned. The sediments yielded unzoned rounded zircons, devoid of crystal faces. The granitised rock contained a small percentage of grains similar to those

TABLE II(a)—SILICATE ANALYSES

	1	2*	3*	4*	5*	6*	7*	8*	9	10	11	12	13	14*	15*	16*	17*
SiO_2	71·40	75·71	75·78	75·06	75·50	76·08	76·16	76·94	71·75	62·26	63·33	69·17	70·26	59·54	57·02	64·43	65·60
Al_2O_3	13·41	11·65	12·46	12·35	11·72	13·35	11·67	12·66	15·20	15·90	16·40	14·59	14·50	17·01	18·37	17·53	18·20
Fe_2O_3	0·62	0·48	0·49	0·94	1·59	0·64	0·64	0·48	0·32	0·99	1·09	0·79	0·64	0·95	0·95	tr.	tr.
FeO	2·67	0·72	1·29	1·15	1·44	0·29	0·86	0·43	1·58	6·17	5·69	2·88	2·01	6·90	7·76	—	—
MgO	0·45	0·39	0·38	0·38	0·41	0·26	0·20	0·29	0·89	3·09	2·90	1·64	1·20	3·77	3·28	0·21	0·09
CaO	2·58	1·55	0·85	0·82	0·96	0·52	0·38	0·61	2·52	1·65	1·87	2·80	2·80	1·23	1·32	3·22	3·13
K_2O	4·41	4·80	3·40	3·50	4·05	3·95	2·70	3·50	3·71	3·60	3·42	3·64	3·53	1·70	1·15	11·33	9·20
Na_2O	3·62	3·95	4·50	4·30	3·70	4·30	5·60	4·00	3·40	3·99	3·30	3·27	3·47	4·55	4·10	2·03	3·20
H_2O-	0·09	0·07	0·24	0·20	0·12	0·08	0·12	—	0·08	0·22	0·22	0·14	0·11	0·12	0·16	0·13	0·08
H_2O+	0·41	0·27	0·34	0·53	0·84	0·55	0·66	0·61	0·56	1·40	1·31	0·72	0·76	3·21	4·06	0·37	0·32
TiO_2	0·40	0·20	0·32	0·31	0·36	0·25	0·25	0·21	0·31	0·70	0·40	0·66	0·50	1·30	1·45	1·00	0·30
P_2O_5	0·19	0·03	0·01	—	0·01	—	—	0·01	0·07	0·25	0·26	0·10	0·09	0·03	0·04	tr.	0·06
MnO	0·04	0·01	0·03	0·03	0·03	0·01	0·03	0·02	0·05	—	—	0·09	0·05	0·10	0·21	0·03	0·01
Cl	—	0·43	0·13	0·09	0·15	0·16	0·50	0·21	—	—	—	—	—	0·20	0·17	—	—
F	—	0·07	0·10	0·06	0·14	0·04	0·05	0·01	—	—	—	—	—	0·02	0·22	—	—
BaO	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	0·21	0·31
SrO	—	—	—	—	—	—	—	—	—	—	—	—	—	—	—	0·09	0·09
CO_2	0·06	—	—	—	—	—	—	—	—	0·04	0·07	—	—	—	—	—	—
TOTAL	100·35	100·33	100·32	99·72	101·02	100·48	99·82	99·98	100·44	100·26	100·26	100·49	99·92	100·63	100·26	100·58	100·41

TABLE II(b)—NIGGLI VALUES

	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15
si	352	450	455	456	450	478	484	504	357	223	235	307	333	207	160
al	38·80	40·30	44·04	44·16	41·82	49·05	43·89	48·82	44·40	33·50	35·80	38·00	40·30	35·01	30·35
fm	16·60	8·50	11·19	12·30	15·27	5·66	8·40	6·69	14·00	37·70	36·80	24·50	19·00	41·30	52·45
c	13·60	10·00	5·41	5·11	5·45	3·04	2·29	4·33	13·30	6·40	7·60	13·30	14·20	4·61	3·88
alk	31·10	41·00	39·35	38·69	37·46	41·89	45·41	40·76	28·00	22·30	19·80	24·20	26·40	19·08	13·32
mg	0·21	0·41	0·29	0·30	0·24	0·40	0·23	0·41	1·00	0·44	0·44	0·45	0·45	0·48	0·62
c/fm	0·82	1·17	0·48	0·42	0·36	0·60	0·27	0·65	0·90	0£17	0·21	0·54	0·75	0·11	0·07
k	0·45	0·44	0·33	0·35	0·42	0·38	0·24	0·36	0·41	0·30	0·40	0·42	0·40	0·20	0·07
ti	—	0·70	1·44	1·49	1·81	1·10	1·14	1·18	1·31	—	—	—	—	3·35	3·30
qz	+128	+186	+198	+201	+200	+210	+202	+244	+115	+34	+56	+111	+222	+31	+17
Shand Symbol	XOmβL	XOmαL	XOmαL	XOmαL	XOmαL	XOmαL	XOmαL	XOmαL							

TABLE II(c)—C.I.P.W. NORMS

	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15
Q	20·06	31·80	33·66	34·70	34·70	34·68	32·52	38·40	29·16	13·14	18·18	26·34	27·84	—	—
or	26·13	28·34	20·02	20·50	23·91	23·35	16·12	20·57	21·68	21·13	20·02	21·13	20·57	—	—
ab	30·39	32·49	38·20	36·15	31·44	36·15	45·06	34·06	28·28	33·54	27·77	27·77	29·34	—	—
an	7·51	—	3·61	3·89	3·34	2·50	—	3·06	12·51	8·06	9·17	13·90	13·90	—	—
c	—	—	—	0·10	—	1·02	—	1·12	1·02	2·55	3·98	0·20	—	—	—
$CaSiO_3$	1·76	2·08	0·23	—	0·35	—	0·70	—	—	—	—	—	—	—	—
di $MgSiO_3$	0·50	1·00	0·10	—	0·20	—	0·20	—	—	—	—	—	—	—	—
$FeSiO_3$	1·45	1·06	0·13	—	0·13	—	0·53	—	—	—	—	—	—	—	—
$MgSiO_3$	0·60	—	0·80	1·00	0·80	0·60	0·30	0·70	2·20	7·70	7·20	4·10	3·00	—	—
hy $FeSiO_3$	2·24	—	1·32	0·92	0·66	—	0·79	0·13	2·11	9·37	9·24	3·56	2·51	—	—
ac	—	0·92	—	—	—	—	1·85	—	—	—	—	—	—	—	—
wo	—	1·16	—	—	—	—	—	—	—	—	—	—	—	—	—
Symbol	1422	1423	1414	1414	1424	1413	1414	1314	—	—	—	—	—	—	—
Q	27·18	31·80	33·66	35·64	35·58	34·68	35·46	39·06	30·72	10·74	19·56	26·70	28·50	14·64	16·86
L	63·75	60·83	61·83	60·64	58·69	63·02	61·18	58·81	64·03	69·38	61·32	62·80	63·81	61·23	57·97
M	8·92	6·77	3·89	3·92	5·22	1·30	3·50	1·99	5·93	19·63	18·79	10·04	7·35	24·50	25·23

1. Porphyritic granite, Vredenburg—D. L. Scholtz.
2. Coarse even-grained granite 1,000 feet W. of Slippers Bay contact—C. E. G. Schutte.
3. Coarse even-grained granite 1,200 feet N. of Slippers Bay contact—C. Lessing, E. C. Haumann.
4. Coarse even-grained granite 75 feet N. of Slippers Bay contact—C. Lessing, E. C. Haumann.
5. Even-grained granite 40 feet N. of Slippers Bay contact—C. Lessing, E. C. Haumann.
6. Even-grained granite 20 feet N. of Slippers Bay contact—C. Lessing, E. C. Haumann.
7. Even-grained granite 7 feet N. of Slippers Bay contact—C. Lessing, E. C. Haumann.
8. Contact microgranite, Slippers Bay—C. Lessing, E. C. Haumann.
9. Hybrid granitised xenolith, Slippers Bay hills—C. J. Liebenberg.
10. Highly contaminated granodiorite, Doornfontein, Darling—D. L. Scholtz.
11. Highly contaminated granodiorite, Klavervlei, Darling—D. L. Scholtz.
12. Partially incorporated xenolith, Klein Paternoster—C. J. Liebenberg.
13. Hybrid microgranite, Great Paternoster—C. J. Liebenberg.
14. Malmesbury hornfels at contact, Slippers Bay—C. Lessing, E. C. Haumann.
15. Malmesbury hornfels 60 feet S. of contact, Slippers Bay—C. Lessing, E. C. Haumann.
16. Microcline-microperthite from porphyritic granite, Slippers Bay—C. E. G. Schutte.
17. Microcline-microperthite from even-grained granite, Slippers Bay—C. E. G. Schutte.

* New analyses

from the granite, some clear unzoned grains and some rounded grains which, if examined in reflected light, showed small irregularly developed crystal faces. Also present were all intermediate stages between the latter two. Zircons from the contaminated portions of the granite were characterised by the presence of markedly elongated commonly zoned crystals in addition to the igneous type.

Among the zircons extracted from the granites at Slippers Bay, the recrystallised contamination and igneous types of M. Wyatt can be readily matched. No intermediate stages between rounded and recrystallised zircons could be observed in the concentrates containing them. This, however, does not necessarily preclude their presence.

It is evident that caution must be exercised in using zircons in petrogenetic interpretations (A. Poldervaart, 1956). In 1950 A. Poldervaart and J. W. von Backström summarised the use of zircon in the granitisation problem as follows: "The granitisation controversy evolves from the production of granite in situ by metasomatism or ultrametamorphism of sediments. If the original sediments contained rounded zircons, statistical studies of zircons of the granitisation series might provide information as to the granitisation process itself. Thus it seems reasonable to suppose that granite formed in situ from such sediments would contain similarly rounded zircons or that the granitisation series would show evidence of recrystallisation of the rounded zircons to sharply terminated prismatic crystals."

Statistical studies of the zircons at Slippers Bay have indicated a decrease in grain-size of the zircons in the granite as the margin of the body is approached and an abrupt transition at the contact from the idiomorphic igneous type zircon to the rounded grains present in the sediments, therefore offering further evidence that the granite in question was not formed in situ.

(d) *Chemistry.*

Eleven new chemical analyses are included in Table II(a). When material was selected for analyses, emphasis was again placed on obtaining samples of both granite and hornfels along a north-south line across the interdigitated portion of the contact. The variations of the Niggli values (Table II(b)) are graphically represented in Fig. 8.

The even-grained granites up to a distance of 1,200 feet from the contact show insignificant variation in chemical composition, a feature which is wholly in accordance with the mineralogical composition of the rocks. The only discernible change in composition is an extremely slight *si* and *al* enrichment and a slight impoverishment of *fm* in the region of the contact. Relative to the central porphyritic granite (No. 1), however, it will be seen that the average composition of the marginal even-grained facies of the granite is distinctly richer in *al, si* and *alk,* thus corroborating the conclusions arrived at by D. L. Scholtz in the case of the hood and core facies of the South-Western plutons of the Cape Province. W. Wahl (1946) believes that the marginal enrichment in *al, alk* and *si* is characteristic of differentiated intrusions and ascribes the variation in composition to thermo-diffusion-convection.

According to C. Boocock, the chemical alteration undergone by the psammitic and pelitic inclusions in the granite of the Cape pluton involves an initial desilication of the xenolith material as a result of the introduction of alkalis

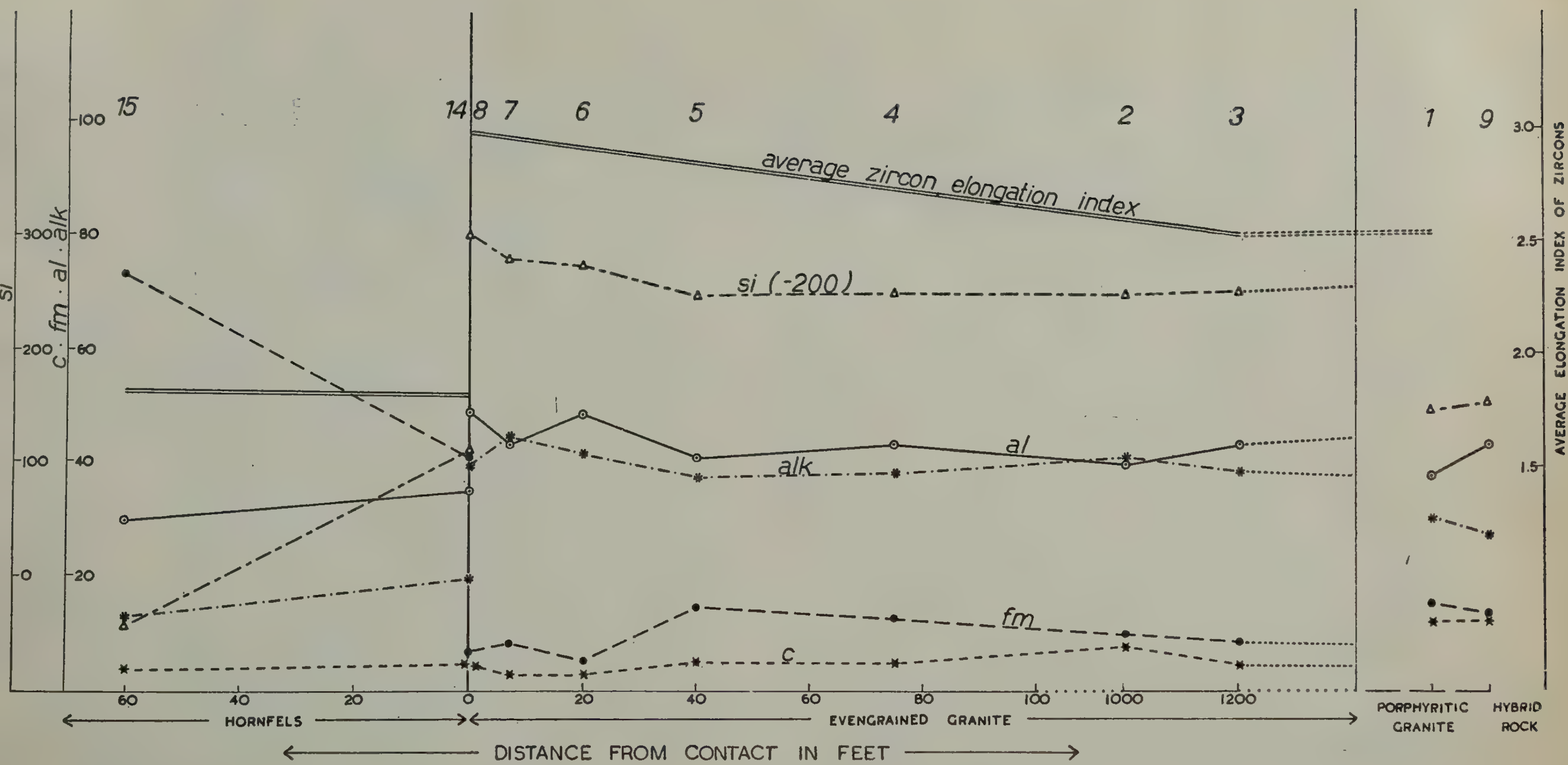


FIG. 8

Variation of Niggli values.

during a process of feldspathisation, followed by the introduction of silica and alkalis during the process of converting the inclusion into granite. These findings therefore accord fully with the views of D. L. Reynolds (1946) in her publication entitled 'The sequence of geochemical changes leading to granitisation'.

By way of comparison the Q. L. M. values of the contact sediments, the granites, contaminated granites and hybrid xenoliths of the Saldanha pluton have also been plotted on the modified Von Wolff diagram utilised by C. Boocock. From the distribution of the added fields there is a still closer agreement with the conclusions arrived at by D. L. Reynolds (see granitisation trend, Fig. 9). Hence there is no reason to doubt that the xenolithic material in both instances reacted with the granite magma in an essentially analogous manner.

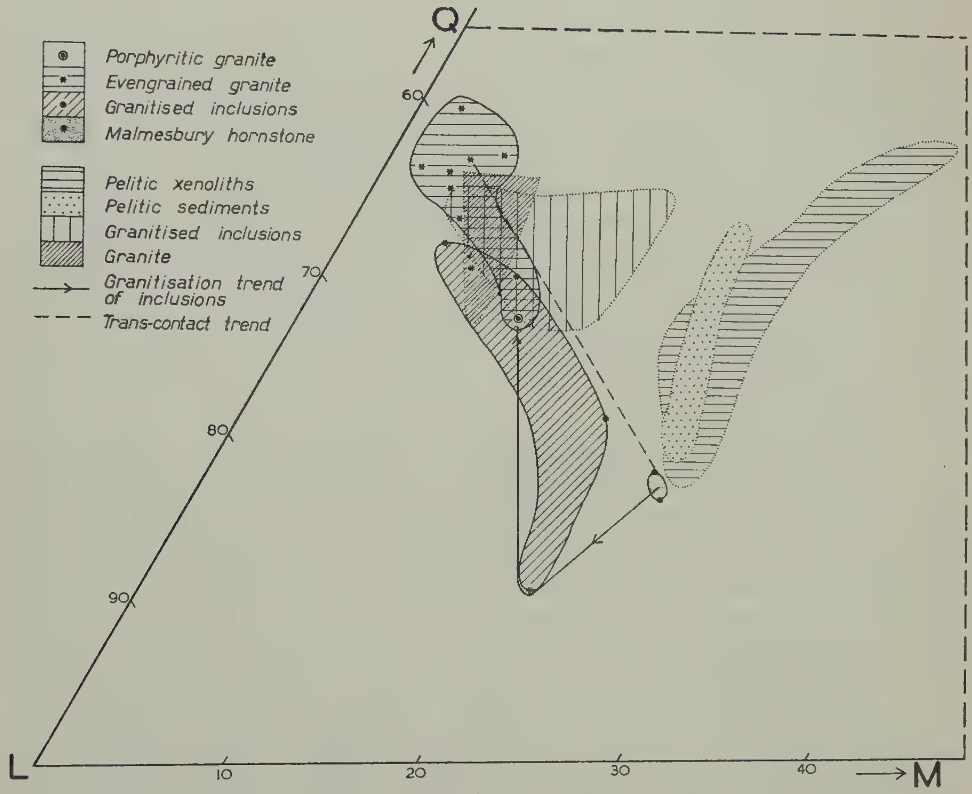


Fig. 9

If we now consider the variation in chemical composition across the contact (see Fig. 8) a very pronounced and abrupt change is immediately apparent. There is not only no suggestion of a gradual transition between the granite and the immediately adjoining hornstone along the contact, but also a complete lack of correspondence in the positions and respective trends

of the *si, al, fm* and *alk* curves of the Niggli values of samples 8 and 14, and 1 and 9. That is, the chemical variation across the contact is different, if not diametrically opposite to that predicted by D. L. Reynolds. In short, we are forced to conclude that there is no evidence whatever of granitisation along the granite-hornstone contact at Slippers Bay, as is also revealed by the transcontact trend on Fig. 9.

From the foregoing it is clear that contrary to the opinion of certain investigators, the critical but cautious use of the variation diagram, in spite of its limitations, still constitutes a valuable petrographic tool which we cannot afford to dispense with at the present stage.

IV. THE BEARING OF THE CONTACT PHENOMENA ON THE CONCEPT OF GRANITISATION

During the past one hundred and forty odd years, numerous scientists of different nationalities have studied varied aspects of the pre-Cambrian granite plutons of the Cape Province. Of these scientists more than thirty geologists have published the results of their findings or criticisms of the conclusions of others not only in international periodicals but also in text-books. J. Otto (1957) has submitted a useful short summary of the literature and conclusions arrived at by the geologists who were particularly interested in the plutons under consideration. It is noteworthy that none of them disputes that the granites bear an intrusive relationship to the invaded Malmesbury sediments. However, differences of opinion about the viscosity of the magma at the time of emplacement are apparent. In this respect the extreme view held by the leading wet transformationist H. H. Read (1951) is significant, since he postulates the invasion of the country rock by a nearly solid and, therefore, presumably highly viscous product derived from the granitisation of crustal material in depth.

In short, all investigators, whether 'pontiffs' or 'soaks' are unanimous as far as the *mobility of the granite as a whole* at the time of emplacement is concerned. Therefore, the only difference of opinion obtains with regard to the magmatic or metasomatic origin of the intrusive *in depth.* This highly speculative aspect of the problem may profitably be disregarded at the present stage in order to focus attention on the wealth of ascertainable facts characterising the accessible parts of the crust.

It is against this background that the contact phenomena at Slippers Bay should be reviewed relative to the concept of granitisation.

When the criteria advanced by the advocates of the transformationist school are considered, the first impression which the area under consideration makes on the observer, is that the Slippers Bay contact provides an excellent example illustrating the phenomena of granitisation in a unique manner. The well exposed and perfectly concordant contact, the conformable attitude of the well-developed cleavage parallel to the bedding of the altered sediments and the faint foliation of the adjoining granite, the presence in the granite of untwinned plagioclase myrmekitically intergrown with quartz, as well as the harmonic relationship between the direction of preferred orientation of spindle-shaped quartz grains and the lineation of the hornstone, appear to be highly suggestive of in situ granitisation. About the last named relation, it should be recalled that D. L. Reynolds (1952), of the extreme dry transformationist

school, regards the similarity in the orientation pattern of quartz grains in many granites and their adjacent sediments as providing "very good evidence for regarding granite as a metamorphic rock".

Closer inspection of the less prominent section of the contact reveals the interlacing and scalloped portion of the contact, recalling the frequently cited in situ "replacement appearance" criterion, which is incidentally also intimately related to the so-called "room problem".

Along this section of the contact the array of aligned lenticular fragments of hornstone, ranging in length from two feet down to microscopic dimensions and in such perfect structural continuity with the neighbouring massive hornstone as if no granite were present, would seem to deliver the *coup de grâce* to the theory of the magmatic origin of granite.

Attention should be drawn to the fact that up to the present stage most, if not all, of the 'so-called' *undoubted criteria* in favour of granitisation as displayed by the Slippers Bay contact, have been enumerated. Let us, however, subject the evidence to closer and more critical scrutiny.

A personal investigation by H. H. Read (1951) of some of the inclusions in the granite led him to remark that from the "great rafts of metasomatised Malmesbury sediments, feldspathised and soaked and veined by granite material; much detail on granitisation can be observed and many valid conclusions may be drawn". D. L. Scholtz (1946) has described the local variations in the composition of the granites owing to contamination: " . . . xenoliths ranging in size from inclusions of microscopic dimensions up to massive blocks of country rock are known to occur well within the granite plutons. In some instances fragments of country rock which had been stoped down to a depth of half a mile or more were mobilised and transformed into dark coloured, often porphyritic rocks, consisting of biotite, hornblende, quartz, scattered phenocrysts of microcline perthite or plagioclase and occasional rounded xenocrysts of quartz. In the more advanced stages of incorporation, the inclusions have been transformed into hornblende and biotite-rich microgranitic material or medium grained leucocratic cordierite bearing granitoid rocks". He concludes "that the possibility of the extensive and complete granitisation of shaly wall and roof rocks appears to be *very remote** in the case of the normal batholithic intrusion, if large down stoped xenoliths or massive blocks of dominantly argillaceous material once completely enveloped by magma under exceptionally favourable conditions still provide mineralogical and/or chemical evidence of contamination". The total absence in the Slippers Bay wall rocks of similar chemical changes as are present in the granitised xenolithic material indicates a complete agreement with the conclusions of D. L. Scholtz.

The orientation of most of the xenoliths (Malmesbury contact, Hoedjies Bay peninsula and elsewhere) not only in relation to the wall rocks, but also to one another shows conclusively that they had at one time been able to move bodily in a yielding medium. In many cases the extremities of the xenoliths show curved, frayed edges which grade imperceptibly into the granite. At Slippers Bay, although the

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*Author's italics.

contact between granite and the wall rock sediments is always sharp, *splaying of the prongs of sedimentary material* in the granite is again in evidence. Now the curvature of some of the otherwise linear parting planes of the sediment could not possibly have occurred in the solid state; and as the original liquidity of the granite is clearly displayed by these evidences of mobility, there can be no doubt that we are dealing with what was once a liquid mass. Further definite proof is afforded by the presence of uninterrupted vein-like off-shoots *from the main granite body* penetrating the country rock, while some idea of the high mobility of the liquid granite is given by the persistence of even the smallest veins of granite in the hornstone at Slippers Bay.

Additional evidence for regarding the granite as the crystallisation product of a melt is afforded by the normal zoning of the plagioclases, which, despite the presence of basic cores, have apparently effectively resisted homogenisation in what appear to be very favourable conditions for granitisation. Support for this view is also forthcoming from the nature of the zircon grains in the granite and adjacent sediments at Slippers Bay. The abrupt transition at the contact from euhedral to rounded grains is reflected in the break in the average elongation index curves (see Fig. 8). The absence at the contact of intermediate stages between the two types of zircon adds further concrete evidence to the ideas favouring a magmatic origin for the granite.

H. H. Read regards the Cape granites as being in the "wrong geological setting for large scale granitisation". This "setting" envisages the aureole of low- to medium-grade thermally metamorphosed Malmesbury sediments, the outer margin of which conforms to that of the granite bodies and fades out away from them. It is impossible to ascribe this aureole to anything but the *heat of the intrusive granite* in which marginal textural variations have been ascribed to the effect of chilling by D. L. Scholtz. This author has indicated the presence of a fine- to medium-grained granite facies which caps the plutons but thins out downwards along the sides of the bodies, and it is below the limit of this facies that belts of migmatite as exposed at Sea Point are developed. Further, the marginal gradation of the "coarsely porphyritic variety of granite into coarsely porphyritic granite porphyry and quartz porphyry by a gradual reduction in the crystallinity of the matrix, the sizes of the phenocrysts remaining approximately constant as in the Stellenbosch and Wellington areas . . . is strongly suggestive of marginal chilling effects". It is noteworthy that the variations of the Niggli values in the core and even-grained granites of the plutons of the Western Province are reflected in the compositions of the marginal and more central portions of the granite along the contact at Slippers Bay. This feature, which W. Wahl (1946) regards as characteristic of differentiated intrusions and attributes to thermo-diffusion-convection, is entirely in agreement with the concept of marginal chilling as described above.

In the Slippers Bay area, despite mineralogical uniformity, there is a decrease in the grain-size of the major constituents and the accessory zircons of the granite as the contact is approached. This decrease is more marked in the vicinity of the interdigitated "replacement appearing" portion of the contact, where the microgranite is developed. The absence of the microgranitic facies along the concordant portion of the contact is to be expected, as the rate of dissipation of heat would have been greater in a direction parallel to the planes of parting than across them. This fact, together with the

conclusions of D. L. Scholtz, leaves no doubt that despite its "replacement appearance" the contact at Slippers Bay *is certainly not* the result of in situ granitisation.

D. L. Scholtz has indicated that there can be no large difference in age between the granite dyke rocks present in the margins of the Saldanha batholith and the batholith itself, as the aplite dykes of the latter can be observed to cut across and include portions of the granitic dykes. These age relationships have been corroborated by J. Otto, who has proved the magmatic origin of the granitic dykes. It would, therefore, be extremely remarkable if the slightly earlier consanguineous granites should have an origin other than magmatic.

If we now consider the criteria which could be advanced by the transformationists, it is evident that these criteria are *swamped by irrefutable evidence indicating the magmatic nature of the granite and consequently their argumentative significance is lost.* With regard to the similarity in the preferred orientation of the quartz grains on either side of the contact, it should be noted that the linear orientation of the quartz in the granite is more perfect than in the adjacent sediments. This does not appear to be indicative of replacement as it could hardly be expected that the derived granite would inherit a more perfect linear structure than that present in the parent rock. D. L. Scholtz considers the preponderence of conformable granite-hornstone contacts characterising the Cape granites as the consequence of the emplacement of the magma along the cores of unsymmetrical anticlines in the Malmesbury sediments, while the reconstruction of events as originally postulated by Charles Darwin in the case of the Sea Point contact, is equally applicable at Slippers Bay and accounts satisfactorily for the orientated xenoliths in the granite in that locality.

V. CONCLUSION

When the evidence for the granitisation origin of the Cape granites is weighed, it is found wanting. One cannot but accept the "replacement appearing" contact at Slippers Bay as a direct result of the emplacement of the granite in a hot, highly mobile state — a conclusion backed by considerable opinion. The contrasting mineralogy of the effusive and intrusive rocks has recently been exploited by the transformationists to justify their conclusions. The presence of the low temperature varieties of feldspar in the Cape granite and the younger acid dykes (J. Otto), however, also casts grave doubt on the validity of the most recent argument advanced by the above mentioned school of thought.

In conclusion the writer would like to quote the words of that distinguished and experienced geologist, Prof. P. Escola, who has spent his life in those areas most frequently referred to when granitisation is discussed: "I believe that the physico-chemical foundation laid by Bowen, Niggli, Goldschmidt and others will last. Many of the advocates of the present day metasomatic school are likely to throw away the baby with the bath-water when wanting to eliminate magma or replace it by something that is beyond the possibility of experiment and observation. But, I also believe that the spirit of this school will result in fair progress of our science when once its *exaggerations** are recognised and avoided."

*Author's italics.

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PLATE I

Ph. 1 Concordant granite-hornstone contact at Slippers Bay.

Ph. 2 Discordant granite-hornstone contact at Slippers Bay.

Ph. 3 Orientated hornstone xenoliths in granite along transgressive contact. (Published by permission of the Geol. Soc. S. Afr.).

Ph. 4 Interdigitated contact showing bending of the prongs of hornstone. (Published by permission of the Geol. Soc. S. Afr.).

Ph. 5 Ptygmatically folded veins of granite in hornstone.

Ph. 6 Granitised inclusion of hornstone in granite.

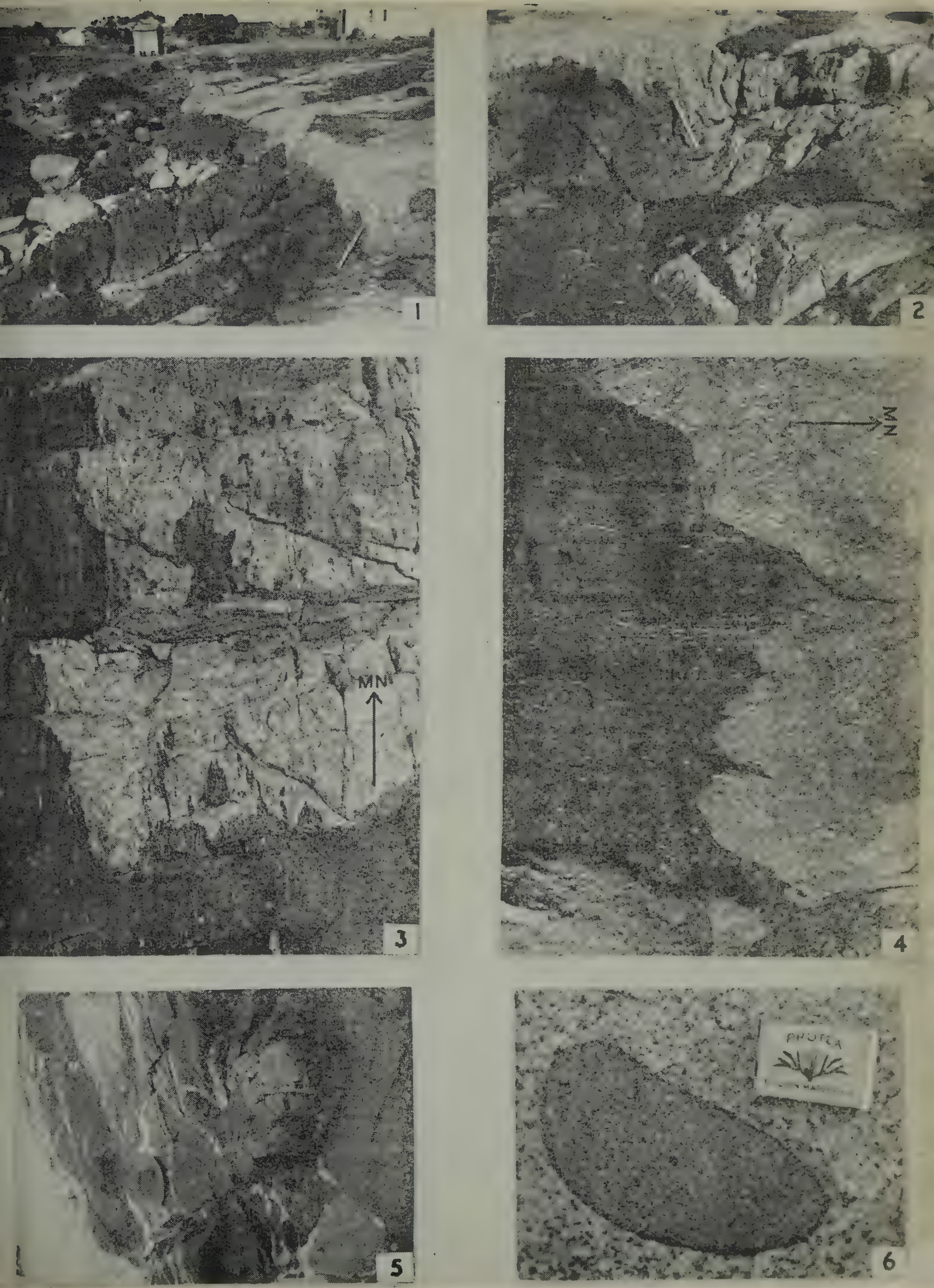

PLATE I

PLATE II

Pm. 1 Sharp contact between hornstone and granite vein. (× 15).

Pm. 2 and 3 Highly zonal plagioclase with labradorite cores. (× 25).

Ph. 4 Granite vein with schlieren of hornfelsic material.

Pm. 5 Semi-opaque zircon. Porphyritic granite. (× 230).

Pm. 6 Zoned elongated zircon. Porphyritic granite. (× 230).

Pm. 7 Zircon with core and zoning. Porphyritic granite. (× 230).

Pm. 8 Clear glassy zircon with bubble inclusion. Porphyritic granite. (× 230).

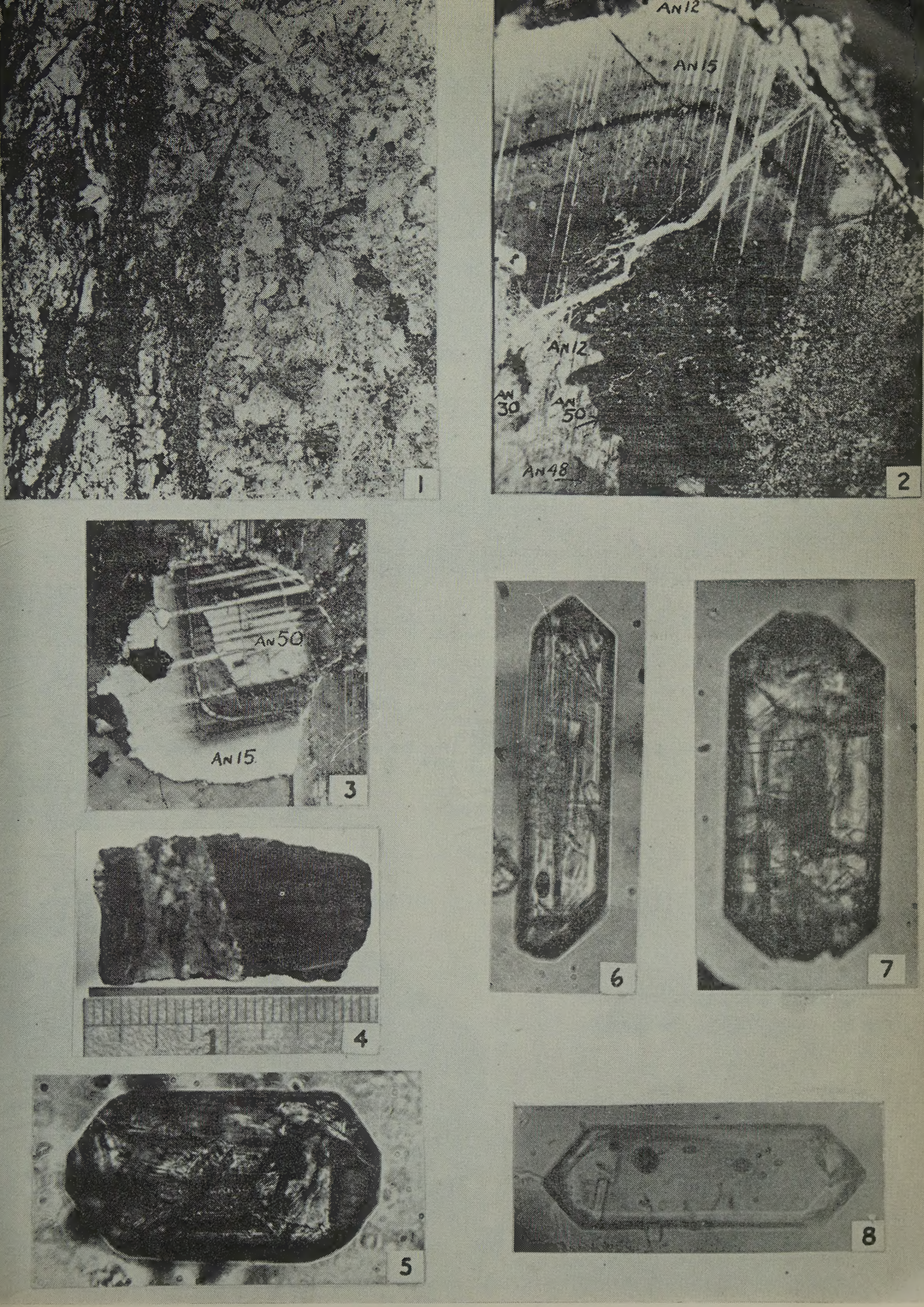

PLATE II

PLATE III

GROUP A.

Pms. 1, 2, 3, 4	Rounded zircons from hornstone. (× 230).

GROUP B.

Pm. 1	Clear elongated zircon. Contact microgranite. (× 230).
Pms. 2, 3	Clear zoned zircons with inclusions. Contact microgranite. (× 230).
Pms. 4, 5	Zoned zircons with cores. Contact microgranite. (×230).

GROUP C.

Pm. 1	Fractured zoned zircon. Medium even-grained granite, 1,200 feet N. of contact. (× 230).
Pm. 2	Unusual stumpy zircon. Medium even-grained granite, 1,200 feet N. of contact. (× 230).
Pm. 3	Semi-opaque zircon. Medium even-grained granite, 1,200 feet N. of contact. (× 230).
Pm. 4	Zoned zircon. Medium even-grained granite, 1,200 feet N. of contact. (× 230).

PLATE III

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